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MINERALOGY AND GENESIS OF THE MAYVILLE IRON ORE OF WISCONSIN

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I. INTRODUCTION AND ACKNOWLEDGMENTS

This paper presents new mineralogical and chemical data on the Mayville oolitic iron ore of Eastern Wisconsin. These, together with a brief discussion of its genesis, are the results of a study begun in 1925 and lately resumed in order to confirm and supplement the earlier work.

The first study was carried out at the University of Wisconsin under the direction of Dr. C. K. Leith. In this the senior author had associated with him Drs. C. H. Stockwell, R. H. B. Jones and E. J. Wechter, the primary purpose of the study then being to ascertain the mineralogical distribution of the phosphorus, since the relatively high content of this element greatly restricted utilization of the ore in the blast furnace at Mayville. This step was essential for a later investigation of methods of beneficiation by reduction of the phosphorus to within Bessemer limits. Grateful acknowledgment is herewith made to Dr. C. K. Leith for suggesting the work, to the Youngstown Sheet and Tube Company, Limited, for samples of the run-of-mine ore with analyses, and to all taking part in the initial investigation. Corroboration of many mineral species noted in the optical studies was given by *x*-ray photographs (powder method) made by Drs. C. O. Swanson and C. H. Stockwell.

The restudy of the ore has been completed at Queen's University by the junior author (A.P.B.) as a thesis problem, with facilities not previously available, which have enabled the more accurate determination of doubtful species and the discovery of some other rare minerals not formerly noted. The present report is thus largely of his writing. For the restudy, ore of a slightly different character was used, samples of crushed ore having been obtained from the U. S. Bureau of Mines Experiment Station, Minneapolis, and uncrushed samples from the Geological Museum, University of Wisconsin.

Table I is a compilation of the minerals found in the various samples of the ore examined, the greater number of which make up only a small percentage of the ore as a whole.

TABLE I. MINERALS IN THE MAYVILLE IRON ORE

	In Oolites	*In Matrix*	
Oxides.	Goethite*	Goethite	
		Hematite	
	Hematite (specular and earthy)	Magnetite	
		Quartz	
	Quartz (as nuclei)	Corundum (rare)	
		Tridymite	(In volcanic fragment)
	Opal	Cristobalite	

Carbonates.		
	Calcite*	Calcite
	Dolomite	Dolomite
	Siderite	Siderite
Silicates.		
	Halloysite*	Orthoclase
	Leverrierite	Microcline
	Newtonite	Anorthoclase
	Chlorite	Perthite
	Neotocite	Plagioclase
		Biotite
		Hornblende
		Chlorite
		Allophane
		Stevensite
Phosphates.		
	Variscite	Collophanite
	Metavariscite	
	Evansite	
	Ceruleolactite	
	Dahllite	
	Collophanite	
	Borickite	
	Voelckerite	
	Koninckite	
	Phosphosiderite	
Others.		
	Five others unidentified	Pyrite
		Pyrrhotite
		Gypsum
		Fluorite
		Northupite
		Wapplerite
		Hydrocarbon

* Most common.

II. OCCURRENCE OF THE MAYVILLE IRON ORE

The Mayville iron ore occurs in several isolated localities in Eastern Wisconsin as lens-like bodies lying unconformably between the Maquoketa (Richmond) shale and the Niagara dolomite. The maximum thickness of ore known is about 55 feet,[1] but at most exposures of the horizon concerned iron ore is either absent or is represented only by a thin ferruginous dolomite layer.

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[1] Thwaites, Frederik T., Recent Discoveries of "Clinton" Iron Ore in Eastern Wisconsin: *U.S.G.S. Bull.* **540,** p. 338.

The age of this ore was long assumed to be Clinton, on account of its lithological and stratigraphic similarity to the Clinton ore of New York. Savage and Ross, however, have shown it to be of Late Maquoketa age,[2] and propose the name *Neda Iron Ore* for the formation.

The ore is a well-stratified deposit of oolitic texture, aptly described by the term "flaxseed ore." Type specimens consist predominantly of hydrous iron oxide and hematite, but calcareous zones, shaly bands and pebbles, and near the top, non-oolitic and pyritic layers are present.[3]

III. PETROGRAPHY AND MINERALOGY

Methods of Study

The texture of the ore and the structure of the oolites is well displayed in both thin and polished sections. Most of the minerals present, however, are masked by the iron oxides, hence their identification was possible only in concentrates prepared by the usual heavy liquid, magnetic[4] elutriation and acid treatment methods of mineral separation. The mineral determinations were made on the basis of optical properties, checked in some cases by *x*-ray diffraction patterns or microchemical tests. The refractive indices and dispersion of several minerals were determined by the Emmons double variation method, but its use was greatly limited by the rarity of some species. A film method of making thin sections,[5] recently described, makes microphotography of the ore easier than is possible with polished sections.

The Oolites

Oolitic Structure. The oolites are small oblate spheroids flattened in the plane of the bedding. Their average maximum diameter is 1 millimeter. They consist of numerous concentric shells of differing mineral and chemical composition. In sections the shells are seen as concentric rings of variable colour and width (Plate A, Fig. 1) in many instances surrounding a central nucleus of organic or detrital material.

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[2] Savage, T. E., and Ross, C. S., The Age of the Iron Ore in Eastern Wisconsin: *Am. Jour. Sci.*, vol. **41,** 1916, pp. 187–193.

[3] Savage, T. E., and Ross, C. S., *op. cit.*

[4] Magnetic concentrates were made with a Dings Magnetic Concentrator, University of Wisconsin.

[5] Appell, J. E., A Film Method for Studying Textures: *Ec. Geol.*, vol. **29,** 1933, pp. 383–388.

Various "abnormal" types of oolites are described under the discussion of Nuclei.

MINERALOGY. A surprisingly large number of minerals occur in the oolites, including a variety of uncommon species. Those present in sufficient quantity to permit their determination are described below. They are referred to those known minerals with which their properties agree most closely, and are listed in order of abundance.

Indices of refraction, where given to the third decimal place are correct to ±0.001.

Goethite ($Fe_2O_3 \cdot H_2O$)

On vertically-illuminated polished sections of the ore the predominance of this mineral gives a dull grey colour to the oolites. Its presence was confirmed by x-ray powder photographs.

Calcite, Dolomite, Siderite

Calcite in the oolites is invisible in thin sections, but was identified by its optical properties in oil immersion mounts of oolitic material. Dolomite is much less abundant than calcite. and siderite was found only in two or three mounts of oolitic material.

Halloysite ($Al_2O_3 \cdot 2SiO_2 \cdot nH_2O$)

Halloysite makes up the bulk of the insoluble residue left after treatment of the oolites with strong HCl and HNO_3. Under a binocular microscope this mineral appears white and finely-granular like unglazed porcelain. It is in fragile spheroidal shells, forming the framework, at it were, of the oolites. Under the microscope the halloysite is white, with irregular fracture and a rough fracture pattern on its surface. It is very difficult to free the mineral completely of iron by the acid treatment; most of the grains retain a pale brownish colour. The intimately-held iron makes the material slightly anisotropic and raises the index of refraction and the dispersion. Besides this impure halloysite, occasional rose to orange-red grains of what appears to be stained halloysite are seen. It was suspected that the colour might be due to manganese, but a spectrographic examination[6] of a small amount of the material revealed the presence of only Al and Si plus minor amounts of Mg and Fe. Optical measurements on the purest halloysite gave $N_D = 1.540 - 1.550$, $N_{(F-C)} = 0.007 - 0.010$. Its determination was verified by x-ray and chemical methods.

Leverrierite ($Al_2O_3 \cdot 3SiO_2 \cdot nH_2O?$)

Leverrierite forms a small part of the insoluble residue. It was positively identified from x-ray photographs. The mineral most closely resembling leverrierite optically gave:

$$N_g = 1.586,\ N_m = 1.585,\ N_p = 1.537.$$
$$N_g - N_p = 0.049.$$

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[6] For this examination the authors are indebted to Mr. G. A. Harcourt at Queen's University.

These figures were obtained by recording the approximate indices of a large number of grains throughout the study. The mineral is present as vermicular aggregates or warped laths, colourless or yellow to brownish. Many grains show incomplete extinction.

Leverrierite and related minerals have a rather wide range of indices. Some grains that otherwise resemble the leverrierite just described have different optical constants. A basal flake showed $2V=0^\circ\pm$, $N_0=N_m=1.515$. This index increased

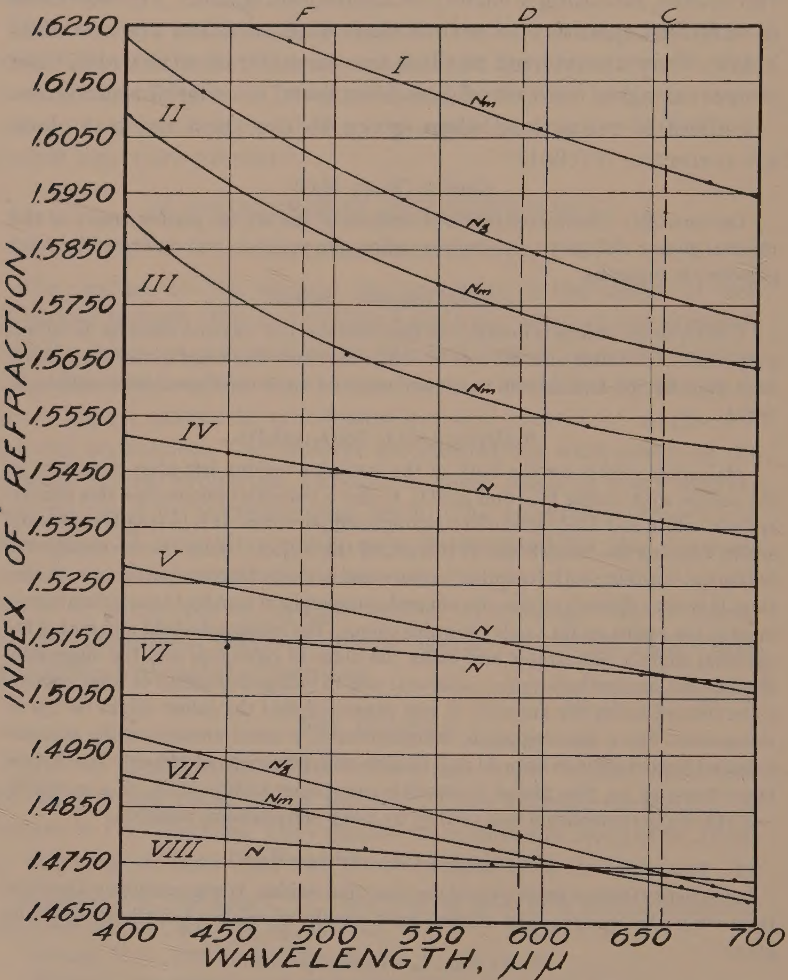

FIG. 1. Chromatic dispersion of some minerals of the Mayville ore. I. Chlorite, II. Variscite, III. Leverrierite, IV. Halloysite, V. Northupite, VI. Stevensite, VII. Tridymite, VIII. Allophanite.

appreciably after several hours immersion in the index liquid.[7] Another lath-like individual gave $N_m = 1.5558$, $N_{(F-C)} = 0.0187$ (Fig. 1).

Variscite ($AlPO_4 \cdot 2H_2O$)

Variscite occurs as small, colourless to very pale pink or yellowish laths with parallel extinction. Measurements of the indices gave:

$$N_g = 1.5857,\ N_m = 1.5737,\ N_p = 1.5570.$$
$$N_g(F-C) = 0.0268.$$
$$N_m(F-C) = 0.0250.$$

Optic sign (+); Z parallel to elongation of laths; X normal to plane of laths. Variscite differs from leverrierite in having lower birefringence, parallel extinction and more regular lath-like grains.

Metavariscite ($AlPO_4 \cdot 2H_2O$)?

Besides the slightly coloured grains of variscite, a pale green pleochroic mineral in similar lath-like fragments is present in smaller amounts. The optical properties determined for it are: Optic sign (−), $N_g = 1.574$, $N_m = 1.556$, $N_p = 1.543$.

$$N_g - N_p = 0.030.$$

Z = light green, Y = pale yellow green, X = colourless. These properties agree best with the data for metavariscite, but are by no means identical with the figures for that mineral.

Opal ($SiO_2 \cdot nH_2O$)

Opal in colourless and tinted grains and fragments of spheroidal shells occurs in the insoluble residue from acid treatment of the oolites. It is isotropic with $N_D = 1.540$.

Dahllite ($7CaO \cdot 2P_2O_5 \cdot CO_2 \cdot \frac{1}{2}H_2O$)

Colourless, fibrous grains and concretionary growths of dahllite are present in some oolites. Elongation of the fibres is negative. Other optical properties are:

Uniaxial negative;

$$N_o = 1.620,\ N_e = 1.605.$$
$$N_o - N_e = 0.015.$$

Phosphosiderite ($2Fe_2O_3 \cdot 2P_2O_5 \cdot 7H_2O$)

The earlier studies suggested this mineral was one of the principal iron phosphates present, though it was recognized that others might be masked by the abundant hydrous iron oxide, Colourless fragments with a perfect cleavage, indices above and below 1.730, and showing strong dispersion, were considered as phosphosiderite rather than siderite, which they resemble somewhat optically, particularly on a basis of the low ferrous iron content of the ore. Chemical tests further showed that fragments of this mineral or related iron phosphates are quantitatively important in some oolites.

Collophanite ($3Ca_3P_2O_5 \cdot Ca(CO_3, F_2, SO_4, O) \cdot nH_2O$)

Blue stained, yellowish and colourless collophanite is present in small amounts. The grains are very irregular in outline and surface and have no cleavage. The optical character of the mineral is isotropic with $N=1.568$. Collophanite was also identified microchemically in the matrix of the ore.

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[7] See Larsen, E.S., *U.S.G.S. Bull.* **679**, p. 193.

Koninckite ($FePO_4 \cdot 3H_2O$)

A pale olive-yellow to greenish mineral without cleavage is optically similar to koninckite as described by Larsen.[8] Most grains are isotropic with $N=1.643$. Other fragments are very slightly anisotropic and have N_m a little greater than 1.644. Phosphorus is present as revealed by a microchemical test.

Borickite ($3CaO \cdot 7Fe_2O_3 \cdot 2P_2O_5 \cdot 24H_2O$)

A wine-red to red-brown isotropic mineral forms some of the oolite shells. It has no cleavage, conchoidal fracture and $N=1.64$–1.67. These properties together with a microchemical test for phosphorus suggest that the mineral is borickite. The latter has a highly variable index of refraction.

Voelckerite ($10CaO \cdot 3P_2O_5$)

Colourless, almost isotropic grains of a mineral resembling voelckerite, with N_g slightly less than 1.6345 and N_p slightly greater than 1.630 were found in the first ore studied.

Evansite ($3Al_2O_3 \cdot P_2O_5 \cdot 18H_2O$)

The mineral identified as evansite lacks colour and cleavage and is isotropic with $N=1.495$. It is insoluble in HCl or HNO_3. Small irregular grains of the mineral have a pearly-white appearance under the microscope.

Ceruleolactite ($3Al_2O_3 \cdot 2P_2O_5 \cdot 10H_2O$)?

Fibrous, lath-like grains of a pale blue colour seem to be uniaxial positive, with $N_e=1.587$, $N_o=1.585$. E(|| elongation) = pale blue, O = very pale blue to colourless. Laths are soluble in HCl since none were seen in the insoluble residue of the oolites. This mineral is very rare in the oolites, but the data obtained strongly suggest that it is ceruleolactite.

Neotocite ($MnSiO_3 \cdot nH_2O$)?

A grain of a light yellow-brown, isotropic mineral with conchoidal fracture was found in one concentrate of oolitic material. Enclosed by the isotropic portion were small, curving fibres of very low birefringence. These had almost the same refractive index as the isotropic part, which was 1.540. The grain was lost before a microchemical test could be attempted, but its physical and optical properties agree closely with those of neotocite.

Besides the minerals described, a number of others are undoubtedly present in the ore in very small amounts, but the data obtained were insufficient to identify them. The approximately-determined optical properties of some of these suggest the presence of newtonite and planerite. At least four other minerals were seen for which the optical character is even less perfectly known.

Of all the minerals described or mentioned above, borickite, metavariscite and ceruleolactitie were not recorded in the original study of the ore. On the other hand newtonite, phosphosiderite and voelckerite were not found during the more recent work. This apparent difference in the mineralogical constitution of the ore is probably related to variation in its chemical composition, as the samples in the two studies were not the same. (See discussion of chemistry of the ore.)

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[8] Larsen, E. S., *op. cit.*, p. 97.

NUCLEI. Most of the oolites in thin and polished section show no central nucleus, but this is in many cases due to the cut of the sections. Four types of nuclei are recognized in the oolites: (a) fossil fragments, (b) mineral or rock fragments, (c) fragments of reworked ore, (d) cruciform nuclei.

The fossil fragments have spherulitic or cellular structures. Most of them resemble pieces of corals or Silurian bryozoa. One coral fragment has been identified by Dr. W. H. Twenhofel as Syringopora. Many of the fossil fragments are replaced by iron oxides, except those that are siliceous. Some are similar to spherulites such as Vaughan describes forming the nuclei of calcareous oolites.[9]

Fragments of quartz, and probably of other minerals, and of rock fragments are not abundant as nuclei in the oolites of the Mayville ore as they are in similar deposits such as the Clinton.[10]

The most important class of nuclei are those consisting of reworked fragments. Among these the following types occur.

1. Angular to rounded fragments of iron oxide.
2. Two smaller oolites forming the nucleus of a larger enclosing one. (See Plate A, fig. 2)
3. Fragment of reworked oolitic ore, containing several oolites, as nucleus of a larger oolite.
4. Fragment of a single oolite as nucleus of a larger enclosing one. In some cases the nuclear oolite is eroded at one end only; at the opposite end the bands are conformable with those of the enclosing oolite.

A fifth type, which is really not an oolite at all, is similar to the "fausse ooliths" described by Cayeux[11] from the French oolitic iron ores. It comprises rounded fragments of reworked ore without enclosing concentric bands. One such grain seen in the Mayville ore consisted of numerous siliceous sponge spicules embedded in iron oxide.

The cruciform nuclei form one of the most interesting features of this ore. As far as can be learned, such nuclei have not been described before from oolitic ores. As the name implies, the cruciform nuclei are small cross-shaped objects. The two arms of the cross are at right angles to one another, and are commonly of unequal lengths. The oolitic shells conform to the outline of the cross for only a short distance outward from the latter. (See Plate A, fig. 3.) One of these crosses, picked from an oolite with a needle, was found to consist of slightly magnetic specular hematite.

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[9] Vaughan, T. W., Oceanography and its relations to other earth sciences: *Jour. Washington Acad. Sci.*, vol. **14,** 1924, p. 327.

[10] Smyth, C. H., *N.Y.S.M. Bull.* **207, 208,** 1919, p. 176.

[11] Cayeux, L., Les Minerais de Fer Oolithiques de France, Fasc. II, Minerais de Fer secondaires, **1922,** pp. 909–911.

THE MATRIX

The non-oolitic portion of the ore consists largely of pore space, by volume. The bulk of the solid material between the oolites is crystalline hematite, with minor amounts of goethite and probably halloysite and leverrierite. Some of the goethite is crystalline with radiating structure. The hematite forms encrustations around the oolites, its crystals projecting into the interstitial spaces and thereby giving the latter the appearance of vug-like cavities. Calcite and dolomite are present in varying amounts with the iron oxides. The small remaining fraction of the matrix contains several minerals in insignificant quantities, as well as a large part of the total organic and detrital matter of the ore, and occasional small phosphatic nodules. Although only single grains of some of the minerals were found, they were often large enough to permit their determination. A number of the minerals mentioned below are of little importance in the study of the ore on account of their rarity, but are considered of sufficient mineralogical interest to merit description.

Northupite, ($MgO \cdot Na_2O \cdot 2CO_2 \cdot NaCl$)

A fragment of white translucent mineral about $\frac{1}{8}$ inch long proved to have properties closely similar to those given for northupite. On one side of this fragment were rounded solution cavities; on the opposite surface a small isometric crystal projected from the main mass of the mineral. The forms present, as noted with a hand lens, were the cube bevelled by narrow faces of the dodecahedron. The mineral is easily soluble in water, and has a rather bitter, saline taste. It effervesces with dilute HCl. Microchemical tests on the aqueous solution showed the presence of Mg, Ca, Na and Cl, the first-named element giving by far the strongest reaction. Calcium may be explained by the presence of minute included grains of an impurity which, from microscopic observation, is probably calcite.

The optical properites of this mineral are as follows:

Colourless; no cleavage. Isotropic, $N = 1.5145$, $N_{(F-C)} = 0.0122$.

The crystal habit, solubility and dispersion given above differ somewhat from the same properties in northupite, as described by Winchell,[12] but otherwise the two minerals are very similar. That occurring in the Mayville ore has possibly a slightly different chemical composition.

Wapplerite ($2CaO \cdot As_2O_5 \cdot 8H_2O$)

A single grain of this mineral was found associated with a green chloritic mineral. The arsenic mineral is white and in powdered form under the microscope shows rhomb-shaped grains with angles of 60° and 120°. On such flakes, which seem parallel to a cleavage, an acute bisectrix figure normal to Z is obtained. The extinction angle $X \wedge$ elongation on such fragments is 21°. The indices of refraction are: $N_g = 1.552$, $N_m = 1.537$, $N_p = 1.528$. Microchemical tests on the mineral gave positive results for both As and Ca and negative for P.

Chlorite (Hydrous Mg, Fe, Al, Silicate)

Several grains of green, flaky material from the matrix apparently belong to the chlorite group. The optical properties of that associated with the wapplerite are as follows:

$(-)2V = 0° \pm$; $N_m = 1.6074 = N_p$ (nearly); $N_m(F-C) = 0.0216$, (see Fig. 1).

$X = Y =$ light green; Z practically normal to the cleavage $\{001\}$.

The mineral is insoluble in cold HCl. Microchemical tests on it show Al and Mg to be present. It is possible grains of this mineral are present also in some oolites.

Stevensite ($H_2Mg_3(SiO_3)_4 \cdot H_2O$)

A small lath-like fragment of a white, waxy mineral proved to have properties close to those given for stevensite. In oil immersion mounts the mineral is colourless with very low relief. It has conchoidal fractures but little or no cleavage. Optically is is isotropic with $N = 1.5113$, $N_{(F-C)} = 0.0053$, (see Fig. 1). The mineral is only slightly soluble in HCl or HNO_3. The solution gave a slight microchemical reaction for Mg.

Allophanite ($Al_2O_3 \cdot SiO_2 \cdot nH_2O$)

A relatively large fragment of colourless mineral, slightly stained, showed a parallel banding of its surface when examined in ordinary light. The grain, however, was isotropic and gave $N = 1.4762$, $N_{(F-C)} = 0.0052$ (see Fig. 1). The banding is considered as possibly derived from the alteration of a plagioclase, or as the imprint of a striated mineral once in contact with this grain. Several other grains without the banding were found during the optical examination of the ore.

The remaining mineral fragments present in small amounts are quartz, feldspars (including orthoclase, microcline, anorthoclase, perthite and plagioclase), bleached biotite, hornblende, magnetite, pyrite, pyrrhotite, fluorite and corundum (sapphire).

The fossil fragments in the matrix are nearly all replaced by hydrous iron oxide. One small piece resembling a brachiopod consisted of gypsum needles with a few grains of calcite.

Another type of organic material in the matrix is a lustrous, black hydrocarbon with conchoidal fracture. It is difficultly fusible, and insoluble in several common organic compounds. This material is probably a variety of mineral coal.

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[12] Winchell, A. N., Elements of Optical Mineralogy, Second Edition, Part II, p. 87, and Part III, p. 178.

The most abundant transported fragments in the ore are rough, angular grains of scoriaceous lava up to $\frac{1}{5}$ inch in diameter. These are largely altered to iron oxide (including much magnetite) but remnants of dirty green to brown glass, dusty with minute inclusions and elongated crystallites, are visible in some of the fragments. The average index of the isotropic glass is 1.53, a value that agrees with those given for glasses of moderately basic composition.[13] Under a binocular microscope some of the lava fragments display flowage lines in well-preserved detail.

One of the freshest-looking of the lava grains, when crushed and examined microscopically, was found to contain two colourless minerals, identified respectively as tridymite and cristobalite. The properties on which these determinations are based are given below.

Tridymite, SiO_2

Tridymite is present as wide laths with narrow interbands of glass or of glass and magnetite. The mineral is colourless with negative relief. The laths appear to be normal to the obtuse bisectrix, and have parallel extinction with $Y \| $elongation, Z normal thereto. Fine parallel lines cross the laths at right angles to the elongation; these are apparently twinning lines, as best shown by insertion of a gypsum plate between crossed nicols. Optical constants are:

$$(-)2V, \text{ moderate (estimated).}$$

$$N_g=1.4805,\ N_m=1.4774,\ N_p=1.4772.$$

$$N_g(F-C)=0.0181.$$

$$N_m(F-C)=0.0132.$$

$$N_g-N_p=0.0033 \text{ (measured directly, } 0.003^{+}\text{).}$$

These figures agree best with those of Mallard[14] measured on a natural prism of tridymite, though N_g and N_g-N_p are higher for the present specimen. The birefringence, however, is closer to that reported in more recent studies of the mineral.

From the above data it appears that the laths of tridymite are sections across basal plates which are separated from each other by films of altered glass.

The grains of tridymite and of the mineral described below together have optical properties similar to those of ptilolite,[15] but are distinguished from the latter by their high fusibility and greater hardness (>ordinary glass). Both minerals, moreover, are unaffected by strong HCl.

Cristobalite, SiO_2

The grains identified as cristobalite are colourless with highly conchoidal fracture, and vary from isotropic to slightly birefringent. Some of the latter show two sets of fine multiple twinning bands intersecting at about 90°. The refractive index of the isotropic form is 1.4858; this figure is the same as that given for the isotropic type, metacristobalite.[16]

Tridymite and cristobalite are of interest in this study since this is the first time, to the writers' knowledge, that they have been described from such sediments.

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[13] George, W. O., *Jour. Geol.*, vol. **32**, p. 365, 1924.

[14] Quoted by Sosman, R. B., The Properties of Silica, **1927**, p. 600; from *Bull. Soc. Fr. Min.*, (1890) vol. **13**, p. 169.

[15] Winchell, A. N., *Op. cit.*, Part II, p. 395.

[16] Winchell, A. N., *Op. cit.*, Part II, p. 55.

Other detrital fragments in the ore consist of well-rounded to sub-angular quartz and quartz-feldspar grains, shaly pebbles, small pieces of friable sandstone and a few thin slips of what seems to be chlorite schist.

Nodules

The small nodules already referred to are roughly spherical in shape, with a maximum diameter of about ¼ inch. When the enclosing iron oxides are rubbed off, the nodules are seen to have a dirty yellowish colour. They have no recognizable internal structure. Mineralogically they are composed chiefly of dolomite and calcite, with a smaller amount of an earthy, yellow mineral. This material was tested microchemically and was found to contain calcium, phosphorus and sulphur. Optically it is isotropic with $N = 1.602$. These properties identify the mineral as collophanite or a similar species.

The nodules also contain minor quantities of one or two other minerals which were not determined. One of these may be koninckite.

IV. CHEMICAL STUDY

Variation in Chemical Composition of the Ore

The following table illustrates the variable chemical constitution of the Mayville ore.

	I	II	III	IV	V	VI	VII
Fe_2O_3	79.25[a]	77.34	76.51	41.67	60.50	79.03[a]	72.34
FeO					0.604		0.44
P_2O_5	1.13	2.75		1.53	3.33	3.05	3.73
SiO_2	6.18	8.57	10.13	15.12	6.90	5.75	5.12
Al_2O_3	2.49	5.00	5.21	12.74	4.72		3.25
MnO		3.50	3.10	2.56	0.20		
CaO	6.81[b]	0.55[b]	0.75[b]	15.48[b]	6.70	5.71	5.98
MgO	0.14[b]	6.64[b]	tr.[b]	8.25[b]	2.97	0.64	0.61
S					0.048		
CO_2					5.86		3.60
H_2O+	4.00[c]	2.65[c]	4.30[c]	2.65[c]	7.06	5.69	4.90
H_2O-					0.95		0.32
Cl						0.003	
Total	100.00	100.00	100.00	100.00	99.842	99.873	100.29

[a] Includes FeO. [b] As carbonate. [c] Water, etc.

I–IV: Analyses compiled by R. D. Salisbury, Geology of Wisconsin, Vol. I, Part II, 1883, p. 305.
V: Composite analysis of run-of-mine ore used in first study, 1925. By E. J. Wechter and Mayville Iron Company.
VI: Partial analysis of hand specimen ore used in later study, 1932–33. A. P. Beavan, analyst.
VII: Partial analysis of sample of crushed ore used in later study. A. P. Beavan, analyst. (Sample from Bureau of Mines Experiment Station, Minneapolis.)

No attempt is made to recast the analyses into minerals on account of the large number of variables involved. It is apparent from the above figures, however, that the constituents vary independently of one another. Some variation in the kind of minerals present as well as in the relative amounts of each might therefore be expected. This appears to be borne out by the different mineral species found in the two ores studied, which show only a relatively slight chemical dissimilarity.

Composition of the Oolites

The composition of the oolites and difference between their composition and that of the ore in which they occur is shown in the following partial analyses (Va and VIIa) of oolites corresponding respectively to ore analyses V and VII above.

	VIIa		Va	
	Total	Soluble	Total	Soluble
Fe_2O_3	69.18			
FeO	0.58			
P_2O_5	4.50	2.62	5.50	5.04–5.20
SiO_2	3.78			
Al_2O_3	4.22			
CaO	6.50	6.50	7.30	7.30
MgO	0.58			0.74–0.78
H_2O+	6.58			
H_2O-	0.38			
CO_2	4.26	4.26	5.15	5.15
Total	100.56			

The analyses show that the oolites are higher in lime and phosphorus than the ore as a whole. In ore VII the oolites constitute

75% of the ore by weight. A simple calculation then shows that the matrix contains 1.42% P_2O_5. The proportion of oolites in ore V was not determined, but chemical tests indicate that its matrix contains an even smaller amount of phosphorus. In both ores the phosphorus is irregularly distributed in the matrix.

CHEMICAL COMBINATION AND DISTRIBUTION OF PHOSPHORUS IN THE OOLITES

Calculations based on the phosphorus minerals and their solubility in various acids indicate that the phosphorus in ore V is combined with ferric oxide, lime and alumina in the proportions 68:26:5.5. The approximate corresponding ratios for ore VII are 45:13:42. The marked difference in the proportions of the phosphates is evident in thin sections of the two phases of the ore: in the one first studied concentric shells of dahllite occur in some of the oolites, whereas in the second type no rings of such transparent material are visible.

The distribution of phosphorus in the ore as a whole has already been mentioned. Within the oolites themselves the element is again not uniformly distributed. The ordinary chemical tests for phosphorus were adapted to staining methods for the detection of the element in the manner described by Holmes[17] for the detection of iron. Much difficulty was encountered, however, in developing a method which would retain the phosphorus compound formed, on polished sections. The most satisfactory results were obtained with a recently published test[18] which, by the use of benzidine hydrochloride, gives a deep blue colouration wherever ammonium phosphomolybdate is formed. Staining by this method shows phosphorus only in certain shells of the oolites. It is almost invariably indicated in their outermost spheroids, a fact previously shown in the earlier work by chemical tests. Many of the inner rings also give the blue stain,but this or other staining methods cannot be used to estimate quantitatively the distribution of phosphorus, since phosphatic minerals are not all equally susceptible to the reactions involved.[19] From the study of ore V above, it is estimated that at least 50% of the total phosphorus in the oolites is contained in their outer shells.

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[17] Holmes, A., *Petrographic Methods and Calculations*, 2nd Impression, p. 269.

[18] Leitmeier, H., and Feigl, F., Ein rascher und empfindlicher Nachweis der Phosphorsaure: *Min. Pet. Mit.*, Band **39**, pp. 224–240.

[19] Leitmeier, H., and Feigl, F., *Op. cit.*, pp. 229–240.

V. GENESIS OF THE MAYVILLE IRON ORE

Origin of Oolitic Iron Ores

A study of the many extensive deposits of oolitic iron ore in the world has lead to the advancement of three theories of origin for those consisting of oolitic limonite[20] and/or hematite. Briefly stated these are:

I. Formation of oolites consisting of various minerals containing iron in the ferrous state, with subsequent alteration under oxidizing conditions to form hydrated ferric oxides or hematite.

II. Deposition of oolitic sediments without important iron content, but later replaced by iron from solutions.

III. Deposition of iron as ferric hydroxide, with later partial dehydration of the iron compound.

At present the first and last conceptions hold most favour among geologists. In the following pages it is endeavoured to show that the Mayville iron ore is of the third type.

The Mayville Ore

Primary Origin. No evidence is known to support the view that the Mayville ore belongs to either types I or II of the above classification. Field relations indicative of important mineralogical changes in the ore are not mentioned in the literature regarding the Mayville deposit. In the laboratory the theory of replacement of an original oolitic limestone would seem to be refuted by the presence of "normal" ferruginous oolites of hydrous ferric oxide in calcareous zones in which the matrix consists almost entirely of carbonate grains with a small amount of interstitial iron oxides. The latter, however, do not replace the carbonates. Consideration of this evidence, in conjunction with that of an even more positive nature given below, leaves no reason for believing that the typical Mayville ore formed in any way other than by direct precipitation of ferric hydroxide and aluminosilicates.

Mode of Deposition. These two substances, ferric hydroxide and hydrated aluminosilicate, are considered to have been deposited from the colloidal state. Apart from the earthy and "amorphous" state of the dominant minerals—a property often, but not necessarily, indicative of colloidal origin—several other features of the ore support the conception of this mode of formation. Chief of these is the concentric structure of the oolites, as held by various authorities.[21] Further, the easy absorption of malachite green by the halloysite of the oolites and the resultant non-pleochroic colour of the mineral is evidence of its amorphous, gel nature according to Grandjean and others.[22] Allophane and opal are likewise mineral gels,[23] and of the phosphate minerals present, variscite, evansite, ceruleolactite, and collophanite have been regarded as of colloidal origin.[24] It is well known that phosphoric acid is adsorbed by many hydrosols, including those of alumina, ferric hydroxide and aluminosilicates.[25]

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[20] Limonite is here used to mean hydrous ferric oxide.

It was formerly thought that calcareous and limonitic oolites grew by accretion of material as they were rolled on the sea or lake bottom. Much experimental and observed evidence has now accumulated, however, which indicates that most, if not all, sedimentary oolites, form by growth in free suspension.[26] In the Mayville ore the presence of reworked material at first suggests the former origin for the oolites, but the following features seem incompatible with such a mode of formation:

1. Oolites lacking a nucleus.
2. Oolites with two unattached oolites as a nucleus.

There is, nevertheless, abundant evidence that all the oolites of the Mayville ore did not originate in situ. The relatively coarse grains of detrital material, the broken oolites, ore grains and shaly particles, the fragmentary character of some of the fossil remains—all are indicative of shallow water sedimentation of a type characterizing deposition and contemporaneous reworking of similar pre-existing material.

On the other hand, the following observations point equally as clearly to the formation of the oolites in situ in an undisturbed environment:

1. Wide range in grain size of the oolites, indicating lack of sorting.

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[21] See Scott, A., The Application of Colloid Chemistry to Mineralogy and Petrology: Fourth Report on Colloid Chemistry and its General and Industrial Applications, *Brit. Assoc. Adv. Sci.*, **1922,** pp. 213–215.

[22] Grandjean, F., *Bull. Soc. fr. Min.*, **1909,** p. 408. Scott, A., *Op. cit.*, pp. 211–212.

[23] Scott, A., *Op. cit.*, pp. 219, 234.

[24] Scott, A., *Op. cit.*, pp. 242–243.

[25] Scott, A., *Op. cit.*, pp. 217–243.

[26] See Bucher, W. H., On Oolites and Spherulites: *Jour. Geol.*, vol. **26,** 1918, pp. 593–609. Pulfrey, W., *Q.J.G.S.*, vol. **LXXXIX,** pp. 416–417. Telfer, L., *Trans. C.I.M.M.*, vol. **XXXVI,** p. 600.

2. Lack of cross-bedding in the ore stratum.[27]
3. Distortion of the banding of one oolite by another in contact with it during growth.
4. Rarely, grains of detrital material (other than nuclei) enclosed in the oolites without distortion of their regular concentric banding.

The obvious conclusion seems to be that both the processes outlined above were active. Thus, while many, perhaps the greater part, of the oolites, formed in situ, others were at the same time washed in from similar deposits previously formed under like conditions of quiet sedimentation.

Source of Material of the Ore. Significant as they are in the foregoing conception of the conditions of sedimentation, the predominant ore minerals throw no light on the baffling problem of their source. In the present study the only information on this question is that yielded by the clastic grains in the ore.

The most abundant transported material, other than reworked oolitic ore and shaly iron oxide, is the fragmental lava. In view of the important part ascribed to igneous activity in the origin of certain iron formations of the Lake Superior region, the volcanic particles in the Mayville ore at once suggest the possibility of Late Ordovician volcanism to which the deposition of the ore might have been related, directly or indirectly. But as yet no igneous activity of such age is recognized in this part of North America. The very minor amount of material volcanic in origin does not justify the postulation of volcanism to explain the genesis of the deposit under discussion, unusual as that type of deposit is among sediments in general.

The alternative explanation of the lava grains is that they were derived from an older land area composed in part, at least, of volcanic flows. The mineralogy of these fragments, described above, suggests that the flows were dominantly intermediate or basic in composition. It is considered that weathering of such lands supplied the highly ferruginous material now forming the Mayville ore.

The thick pre-Niagaran limestones and dolomites of Wisconsin, all of which contain small amounts of iron and alumina,[28] would also appear to offer a possible source for a ferruginous, aluminous sediment free from important quantities of detrital material. Again, siliceous and calcareous iron formations, or pre-existing iron ores, may be suggested as the source of the Mayville ore. Neither of these possibilities, however, is supported by evidence of a positive nature; they are, therefore, not considered further.

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[27] Savage, T. E., and Ross, C. S., *Op. cit.*, p. 192.

[28] Steidtmann, E., Limestones and Marls of Wisconsin: *Wisconsin Geol. and Nat. Hist. Surv., Bull.* **66,** 1924, pp. 183-187.

Assuming, then, that the ancient land area yielding sediments consisted partly of volcanic flows, and further, that no such rocks were formed between Pre-Cambrian and Silurian times, it would appear that over much of this land mass pre-Cambrian rocks were exposed. The present distribution of formations in Wisconsin and adjoining states indicates that the ancient continent, during Late Maquoketa time, lay west and northwest of the area in which the Mayville ore now lies.

These views as to the character and location of the land mass are confirmed by comparison of the detrital material found in the ore with the known types of pre-Cambrian rocks in Wisconsin. The clastic fragments are listed below in order of abundance, together with corresponding rocks from which they were possibly derived.

Basic lava grains....................	Basic flows now exposed south of Lake Superior.
Grains of quartz, feldspars (or both) hornblende and mica..............	Pre-Cambrian igneous rocks of Wisconsin.
Grains of chlorite schist..............	Crystalline schists of Wisconsin pre-Cambrian Rocks.
Grains of sapphire corundum.........	Feldspathoid-bearing rocks in central Wisconsin.
Grains of friable sandstone with well-rounded quartz....................	Pre-Silurian sandstones of Wisconsin.

NATURE OF WEATHERING. Evidence has already been cited, in the reworked character of fragments in the ore, to show that the Mayville deposit was laid down in fairly shallow water. The paucity of clastic fragments and the abundance of meta-colloids in such a formation must therefore bear mutual attestation of chemical decay as the important erosive process, a process promoted undoubtedly by the warm, equable climate prevailing in Late Maquoketa time[29] over a land of low or slight relief and sluggish streams.

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[29] Schuchert, C. A., *A Text Book of Geology*, Part **II**, pp. 232–442.

Secondary Mineral Changes. Neither thin nor polished sections of the ore yield much information as to the association of its minerals, on account of the mantle of opaque iron oxides.

The coating of crystalline hematite surrounding the oolites suggests that this mineral has formed through secondary reorganization of iron oxides since consolidation. The large volume of voids between the oolites may represent a concomitant removal of carbonate matrix, since similar hematite coatings are present on some oolites seen in a calcareous band of the ore. Goethite, it would seem, may be either primary or secondary by hydration of hematite, since ferric hydroxide, precipitated in sea water, tends to take the anhydrous form.[30] The oolitic character and predominance of goethite, however, favor a primary origin for this mineral.

Mineral relations observed under a binocular microscope are:

Association of pyrite with a hydrocarbon.
Encrustations of yellow collophanite on the iron oxide minerals.
Encrustations of carbonates on the iron oxide minerals.

The presence of sulphate-bearing minerals (collophanite and gypsum), secondary carbonates and phosphatic nodules all suggest that circulating solutions have influenced the mineral composition of the ore to some extent. Sulphate minerals were formed possibly by such solutions, deriving that radical from the oxidation of pyrite in the ore and acting on primary constituents of the deposit. The consistently phosphorus-rich outer shells of the oolites, a feature that seems difficult to explain as primary, may be due to solution of phosphorus from the matrix and redeposition on the oolites, or to absorption of phosphatic solutions arising in the matrix, by the gel minerals of the spheroids. There is nothing to indicate introduction of phosphorus from outside the ore bed. The rocks adjacent to the ore horizon are themselves almost devoid of phosphorus.[31]

VI. ECONOMIC ASPECTS OF THE STUDY

As noted in the introduction, the mineralogical study of the Mayville ore was undertaken to ascertain the nature and distribution of the phosphorus minerals preliminary to a study of possible methods of eliminating some or all of these.

The determination that the phosphorus is present as minute

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[30] Lindgren, W., *Mineral Deposits* **(1928)**, p. 313.

[31] Steidtmann, E., *Op. cit.*, pp. 44, 188.

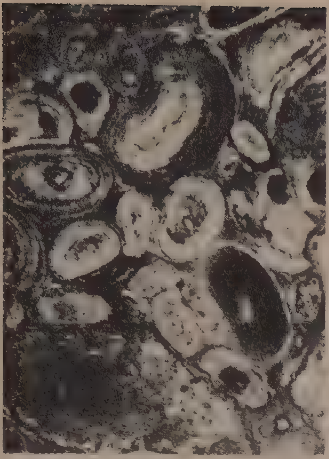

FIG 1

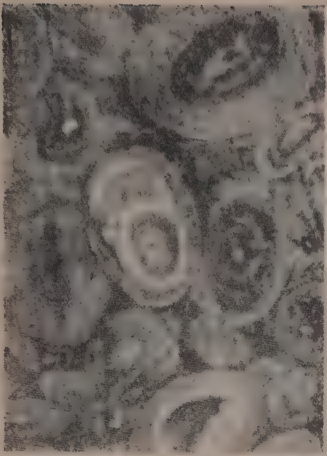

FIG 2

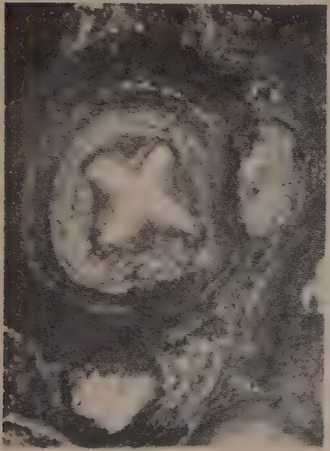

FIG. 3

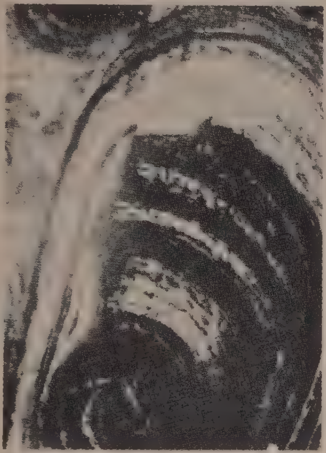

FIG. 4

EXPLANATION OF PLATE A

FIG. 1. Section of Ore approximately parallel to bedding. Note indentation of one oolite by another, with distortion of banding in the larger. A large "false oolite" of iron oxide occupies the lower left-hand corner. ×27.

FIG. 2. Oolite showing a nucleus consisting of two unattached smaller oolites. ×27.

FIG. 3. Cruciform nucleus. A smaller oolite indents the larger with distortion of its banding. ×84.

FIG. 4. Oolite with nucleus consisting of a broken preexisting oolite. The latter itself has also a nucleus of broken oolite. ×84.

particles of iron, lime and aluminum phosphates, intimately associated and intergrown with iron oxides in the oolites, though concentrated to a major extent in their hard outer shells, suggested that no practical method of effecting a separation of these minerals would be found. Nevertheless some experiments were carried out along lines suggested by the earlier work. In this, the fact that most of the ore being used in the blast furnace in 1925 required nodulizing was taken into consideration. The results of certain experiments are summarized below.

1. Acid treatment of raw and nodulized ore. This method, suggested by microchemical tests on oolites, involved the leaching of ore with nitric and hydrochloric acids of varying concentrations. While 94 per cent of the phosphorus could be removed from the raw ore, and material of Bessemer grade produced from the nodulized product, costs of acid were alone prohibitive.

2. Dephosphorization of raw ore by roasting with sodium carbonate and forming water-soluble phosphate gave interesting results but was not sufficient to make the process appear promising.

3. On account of the intimate association of phosphorus minerals with the iron oxides and slight differences in their magnetic susceptibility (particularly of the iron phosphates and oxides) a magnetic separation of the iron ore minerals from those containing phosphorus was not feasible.

4. The low fusibility of the phosphorus minerals renders any magnetic separation after nodulizing and partial reduction of the iron to magnetite impractical, since the phosphorus-bearing constituents become even more intimately locked up with the iron minerals in the sintered product than in the ore.

The results thus confirm the opinion that elimination of the phosphorus can only be obtained economically by metallurgical methods.

OBSERVATIONS ON SPHERULITES

R. J. COLONY[1] AND A. D. HOWARD[2]

INTRODUCTION

Petrographic studies of rhyolitic rocks from Yellowstone Park, Wyoming; Mono Craters, California; and the Lipari Islands, have revealed additional features of interest concerning spherulites. Most of the specimens examined were selected from a collection of Yellowstone rocks made by the junior author in connection with a study of the problem of the history of Yellowstone Canyon, an investigation sponsored by the Princeton University Research Organization. The remainder of the specimens were selected from the collections at Columbia and New York Universities.

CURVATURE OF LINES OF MICROLITES AROUND SPHERULITES

The distribution of microlites in curved lines around spherulites is of interest in view of the generally accepted idea that spherulites form after the lava has come to rest. This feature was observed in the rhyolitic vitrophyre which outcrops immediately back of the rim of the Lower Falls of Yellowstone River. The rock consists largely of glass within which occur spherulites, scattered phenocrysts of quartz and sanidine, and abundant microlites which are strung out in flow lines. The section shows excellent perlitic structure. The perlitic cracks often traverse spherulites, indicating that the spherulites formed before the final contraction of the lava. The spherulites average slightly less than 0.1 mm. in size, and although too small and compact to show distinctly the individual crystal fibers, do show a clear black cross between crossed nicols. The spherulites, with few exceptions, contain scattered microlites which are always in haphazard arrangement within the spherulites themselves, although outside of them they show characteristic alignment. Commonly the lines of microlites bend around the spherulites. This feature is illustrated in Fig. 1.

This phenomenon admits of two possible explanations:

1. The fibers of the spherulites, growing in the hot, plastic glass following the extravasation of the lava, pushed aside or bowed out the streams of microlites in their path.

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[1] Columbia University.

[2] New York University.

2. Renewed local movements in the lava subsequent to the formation of the spherulites swirled the microlites around them.

To determine whether growing fibers in a spherulite can exert an outward push on microlites in their path, the following experiment was conducted. Crystals of menthol were heated on a glass slide and finely comminuted particles of pyroxene were sprinkled into the melt. Because of its acicular habit many of the pyroxene fragments closely resembled in shape the actual microlites of the vitrophyre. The fragments ranged in size from the lowest limits of visibility to 0.5 mm. The cooling process was observed under the microscope. It was found that as long as the melt remained sufficiently liquid the "microlites" were borne toward the growing spherulite with appreciable velocity by convection currents. As the "microlites" were met by the advancing spherulite front they

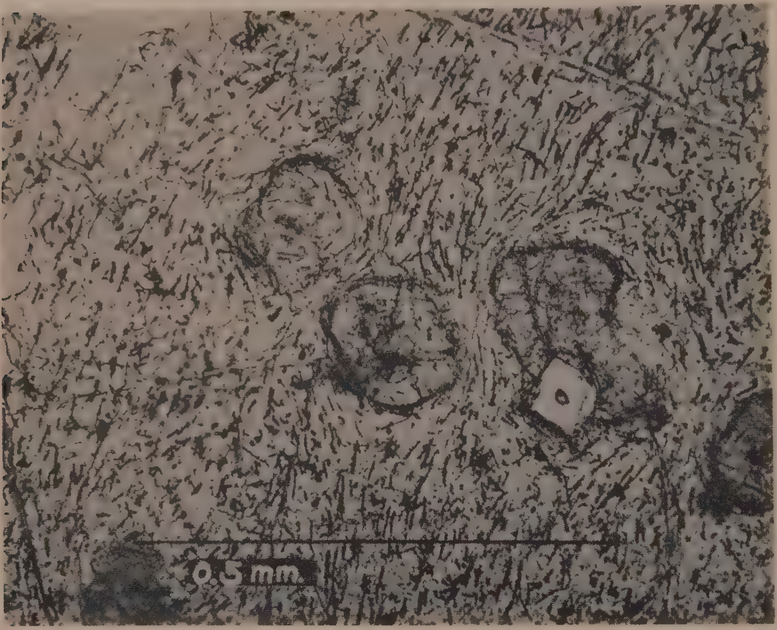

Fig. 1 (Specimen 78). Obsidian just back of the brink of the Lower Falls of the Yellowstone River, Yellowstone Park.

Microlites in haphazard arrangement within the spherulites, but arranged in flow lines without and around the spherulites.

were enveloped by it, and their arrangement within the spherulite was haphazard. As the viscosity of the melt increased the movement of the "microlites" became slower. When the viscosity was such as to prevent further movement of the "microlites" still outside the spherulite, they in turn were overtaken and included within the growing spherulite. In all the various stages of viscosity observed movement of the "microlites" was either toward the spherulite or lacking. Provided the behavior of the acicular pyroxene fragments in the menthol melt is accepted as a sufficiently close analogy to the behavior of microlites in lava, it is judged to indicate that growing spherulites exert no push on obstacles in their path.

The same conclusion is inevitable from the theoretical viewpoint. It is apparent that the continuous growth of a spherulite depends on diffusion of substance from the surrounding magma toward the growing spherulite, and that any factor affecting the rate of diffusion will affect the rate of growth of the spherulite. Furthermore, any factor affecting the rate of diffusion toward a portion of a spherulite, or toward one fiber in a spherulite, will affect the rate of growth of that portion or fiber. An obstacle, such as a microlite or phenocryst, close to the growing front of a spherulite, may affect the growth of that portion of the spherulite by interfering with the diffusion currents supplying that portion with substance. It seems logical to suppose that as long as the obstacle and the spherulite are widely separated the rate of diffusion toward the spherulite will be uniform. This results in uniform growth. When, however, the spherulite front has advanced close to the obstacle, the fibers immediately to the leeward of the obstacle will obtain their supply of substance from currents sweeping around the obstruction. Since their source of supply is now no longer directly ahead of them, but to either side, further growth of these fibers will take place in the new directions. The ultimate result will be that the fibers will curve around the obstacle in their path. The curvature, however, since initiated before the growing fibers reached the obstruction, *is not due to actual impact with the obstacle.*

The third objection to the hypothesis of "push action" is the fact that the lines of microlites do not curve around all the spherulites. Why this "push action," provided there is such an action, should be selective is not clear, especially since the uniformity in size of the spherulites would seem to indicate that the spherulites all grew at about the same time, and since the haphazard arrangement of the included microlites seems to indicate that the lava, at the time of spherulite growth, was fairly liquid.

The alternative is that the curvature of the lines of microlites is a result of renewed local movements in the lava. That this feature is found in connection with only a portion of the spherulites would seem to indicate either of two histories:

1. After the lava came to rest, but while still more or less liquid, a first generation of spherulites was formed. These were caught up in slightly later local movements of the lava with the result that streams of microlites were drawn around them. Following these secondary movements a second generation of spherulites was formed.

2. After the lava came to rest *all* the spherulites were formed. Somewhat later, while the lava was still more or less liquid, small scale, perhaps even microscopic, movement of local character caused the swirling of microlites around *some* of the spherulites. Such localized movements may be due to convection currents generated as a result of variations in viscosity, composition, density, temperature, pressure, or to a "settling" of the lava body.

The uniformity in size of the spherulites in the sections examined is considered unfavorable to the first possibility because it seems improbable that the second generation of spherulites attained just the size of the first when growth stopped. In the writers' opinion, therefore, the curvature of the lines of microlites around the spherulites is due to renewed movements in the lava, local and small scale in character subsequent to the cessation of spherulite growth.

SPHERULITES WITHIN SPHERULITES

The western rim of the Grand Canyon of the Yellowstone about a half mile north of the Lower Falls is composed of a dense, porcelain-like, white, rhyolitic rock which is traversed by an intricate network of tiny fractures. Embedded in the porcelain-like matrix are occasional quartz phenocrysts rarely exceeding 2.0 mm. in size. In thin section the rock is seen to include within a complex matrix of glass, opal and microcrystalline silica, phenocrysts of quartz, abundant spherulites and an undetermined substance[3] which is often pseudomorphous after feldspar euhedra (Fig. 4). The spherulites occur up to 8.0 mm. in size although averaging only about 0.1 mm. The larger spherulites often contain within themselves smaller spherulites, as well as occasional phenocrysts.

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[3] This substance is colorless, cryptocrystalline to almost isotropic, with an index

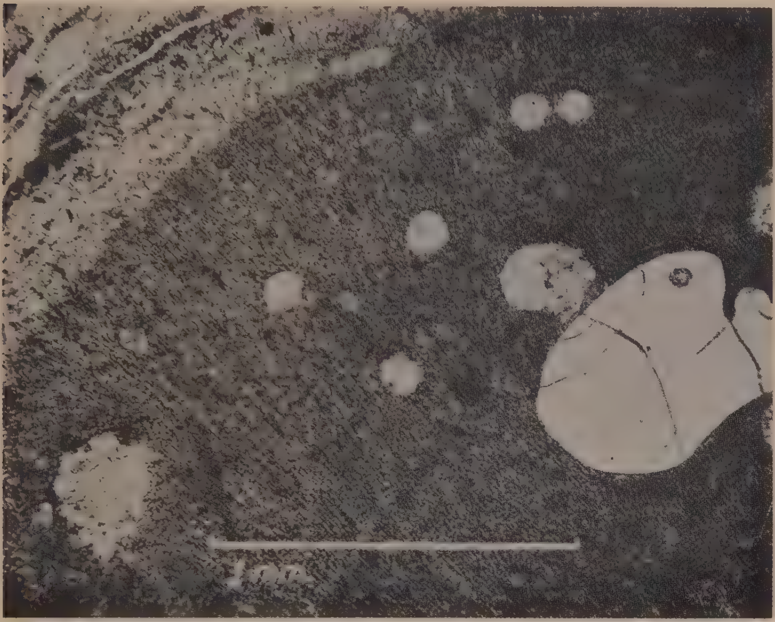

FIG. 2 (Specimen 93). Specimen from the western rim of the Grand Canyon of the Yellowstone at the head of Red Rock trail, Yellowstone Park.

Spherulites within spherulites. A number of small spherulites and a corroded quartz crystal are included in a large spherulite. The crystal fibers of the large host spherulite curve around the small included spherulite shown in the left lower corner of the photomicrograph.

The phenomenon of inclusions within spherulites has been noticed many times. Iddings,[4] in discussing the compound spherulites of Silver Cliff, Colorado, says "the plumose rays shut in as inclusions between them much magma that solidified in other ways." More recently Foshag[5] described spherulites "made up entirely of rod-

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between those of Canada balsam and quartz. It commonly contains threads and filaments of a component with a different index which polarizes differently. It is at present under investigation.

[4] Iddings, J. P., Igneous Rocks: I, p. 240, 1909.

like feldspar crystals and tridymite plates with occasional intratelluric feldspar and quartz phenocrysts embedded within their mass"; and further, "These pherulites appear to be the result of a progressing crystallizing process that started from a nucleus and spread outward through the glass and engulfed some of the phenocrysts." The phenomenon of "spherulite inclusions" within spherulites, (Fig. 2,) has never been described so far as the writers know. An understanding of the conditions productive of this phenomenon may shed light on some of the changes which take place in cooling rhyolitic lavas.

The spherulites, both large and small, are similar in every respect but size. It has been noticed that the crystal fibers of the host spherulite curve around the included phenocryst or spherulite. The curvature of the crystal fibers admit of two possible explanations:

1. The included spherulite is later than the host spherulite; in its growth it pushed aside the fibers of the host spherulite.
2. The included spherulite was captured by the host spherulite as the latter grew outward into the surrounding magma. The fibers of the later, host spherulite, on meeting the spherulite in their path, were deflected around it.

If the first explanation is true then we are forced to believe that the growing fibers of the included spherulite were able to push aside the previously formed fibers of the host spherulite. The objections to any such pushing action have already been discussed. A further objection arises when one considers that the included spherulite, if it is assumed to have formed later, must have drawn its supply of substance from its immediate surroundings; that is, from the host spherulite. Yet the host spherulite must have already used up all the substance necessary for the growth of its own fibers. It would appear, therefore, that the included spherulites are not later than the host, but had reached their present size before being overtaken and included within the host spherulite. The inclusion of the smaller spherulite may have occurred, conceivably, in any one of three ways:

1. Two spherulites may have started growth from two closely spaced centers at the same time. For some reason one ceased growing. The spherulite which continued to grow finally included its neighbor.

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[5] Foshag, W. F., The minerals of Obsidian Cliff, Yellowstone National Park, and their origin: *Proc., U. S. Nat. Mus.*, vol. **68**, Art. 17, p. 6, 1926.

2. The smaller spherulite may have grown to completion first. Later crystallization from a nearby nucleus may have resulted in a second spherulite which grew to large size, including the earlier formed spherulite.

3. The larger spherulite may have started to grow first. Later, when the first spherulite had reached a fairly large size, crystallization from a nearby nucleus started. Continued growth of the larger spherulite resulted in the inclusion of the smaller.

With reference to the first possibility, it is considered unlikely that of two spherulites growing within a few millimeters of each other and which started to grow at the same time, that one should continue to grow and the other stop. Such a condition would require that the magma surrounding the spherulites was unable to further contribute substance to one spherulite, which, therefore, ceased growing, but was able to contribute to the growth of the second spherulite as the growing front of the latter reached this area. The same objection holds for the second possibility, where the smaller spherulite is assumed to have grown first, and later crystallization from a nearby nucleus to have resulted in a larger spherulite, which, as it grew outward, included the earlier formed spherulite. Furthermore, it is not clear why the renewed crystallization, following cessation of growth of the first formed spherulite, should start at a new locus only a few millimeters away, rather than at the locus already provided.

The probable answer appears to lie in the last alternative. According to this idea the larger spherulite is assumed to have grown to appreciable size before the initiation of growth of the smaller spherulite nearby. As it grew outward the larger spherulite included the smaller even though both may have been growing at the same rate. As soon as the smaller spherulite became included in the larger its growth stopped because of lack of substance, while the host spherulite continued to grow so long as diffusion from the surrounding magma continued. This conception, involving initiation of spherulitic growths in the magma over a period of time, can be taken as additional evidence of the heterogeneity of lava. It is somewhat more satisfactory than a concept involving simultaneous initiation of spherulite growth throughout the mass, as could obtain only in the case of a perfectly homogeneous lava under perfectly uniform conditions of cooling.

SIGNIFICANCE OF THE ARRANGEMENT OF MICROLITES IN SPHERULITES

If the manner of inclusion of "microlites" in artificial menthol spherulites is similar to that which occurs in glassy, volcanic lavas, then an additional point of interest arises.

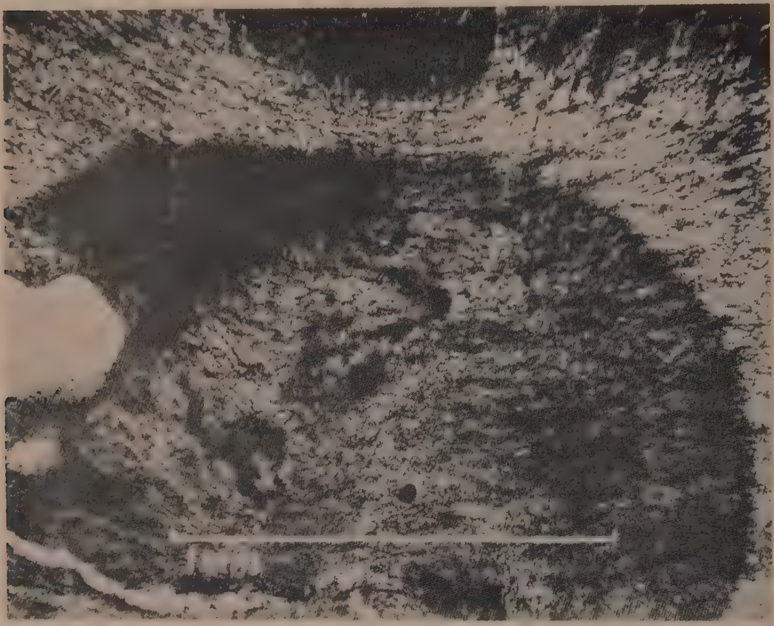

FIG. 3 (Specimen 234). Specimen from the west side of Gardiner River one mile north of Seven Mile Bridge, Yellowstone Park.

Microlites passing through a large spherulite in undisturbed flow lines.

It has been shown that so long as the menthol remains quite liquid the microlites move toward the growing spherulite and are incorporated in it in haphazard arrangement. When the menthol becomes highly viscous, however, there is no movement of the microlites and they are incorporated in the spherulite in the arrangement they had when viscosity overtook them. The arrangement of microlites in spherulites is, therefore, an index of the viscosity of the lava at the time of spherulite growth, i.e. immediately after the lava came to rest. In making use of this criterion it is essential that a sufficient number of thin sections be available so that an accurate cross section of the character of the lava is obtained. It is realized, in view of the heterogeneity of lava, that certain portions of the flow may cool more rapidly than others, so that in some portions of the flow the microlites may occur in flow lines in the spherulites, while elsewhere they may occur in haphazard arrangement.

When the microlites are included within a spherulite in undisturbed flow lines, as is the case in the rhyolitic lavas at several localities in the Yellowstone (Fig. 3), in many of the obsidians from Mono Basin, California, and in some of the lavas of the Lipari Islands, the inference is that at the time of emplacement the lava was so viscous that the microlites were unaffected by diffusion toward the centers of spherulite growth. On the other hand, when microlites are included within spherulites in haphazard arrange-

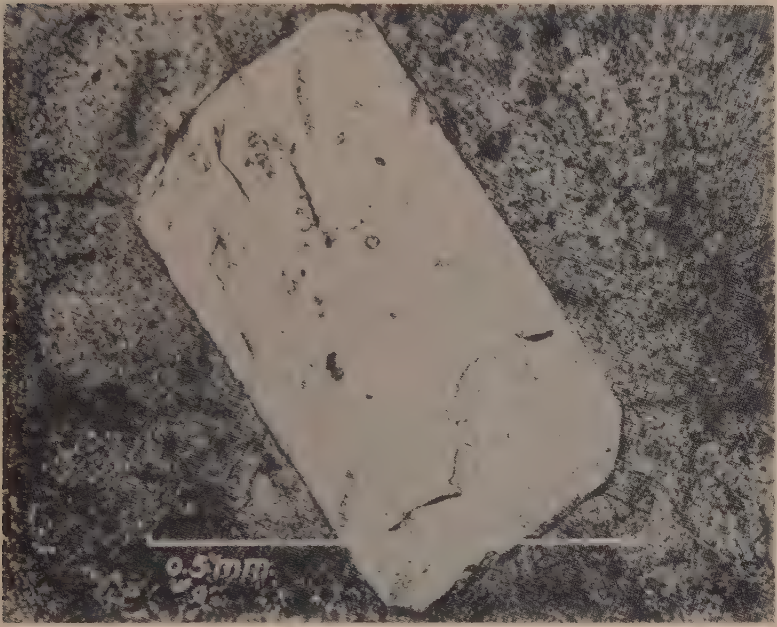

FIG. 4 (Specimen 93). From the western rim of the Grand Canyon of the Yellowstone at the head of Red Rock Trail, Yellowstone Park.

The undetermined substance, pseudomorphous after an idiomorphic feldspar crystal, showing the minute filaments of a second substance with a different index. The mass as a whole is almost isotropic.

ment, while arranged in flow lines outside the spherulites, the inference is that at the time of emplacement the lava was sufficiently liquid so that the microlites were disturbed by the diffusion currents set up in reponse to spherulite growth. This condition obtains in the rhyolitic vitrophyre back of the Lower Falls of the Yellowstone, and is shown in Fig. 1.

Because of the close relationship between viscosity and the rate of cooling the arrangement of the microlites in the spherulites may be used as an indication of the comparative rates of cooling of similar lavas.

THE UNMIXING OF CHALCOPYRITE FROM SPHALERITE

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Previous Work

Schneiderhöhn and Ramdohr[1] suggest that certain intergrowths of sphalerite and chalcopyrite may be due to an unmixing of chalcopyrite from sphalerite. To the best of the writer's knowledge, Schwartz[2] is the only one who has carried on any experiments which might show chalcopyrite dissolving in sphalerite. Nine specimens from Leadville, Colorado, were heated at temperatures ranging from 200° to 650°C. for time intervals of from 22 hours to 240 hours. The results of all runs were negative.

Gross[3] states that chalcopyrite is oriented parallel to the sphalerite (100) planes and Gruner[4] says that it is just as commonly oriented parallel to the (111) sphalerite planes.

Professor W. H. Newhouse, in an examination of some specimens of sphalerite from Bingham Canyon, Utah, noted blebs of chalcopyrite arranged in planes within the sphalerite. These small masses of chalcopyrite were observed near or on the borders of clear growth bands in the sphalerite. A specimen was heated for one week and then quenched in water. Dark haloes seemed to have developed about the chalcopyrite blebs. This strongly suggested that the chalcopyrite had dissolved in the sphalerite. These observations led to the present investigation of the phenomenon.

Method of Attack

The writer has been fortunate in obtaining euhedral sphalerite from Bingham Canyon, Utah. The larger crystals in thin section show growth zones ranging from clear and transparent to opaque. The chalcopyrite blebs seem to be confined to the clear zones adjacent to the opaque zones, and probably merge into the opaque zones.

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[1] Schneiderhöhn, H., and Ramdohr, P., Lehrbuch der Erzmikroskopie, Bd. II, *Berlin*, **1931**,, pp. 106–110.

[2] Schwartz, G. M., Experiments with sphalerite-chalcopyrite intergrowths: *Bull. Geol. Soc. Amer.*, vol. **42**, 1931, pp. 187–8.

[3] Gross, R., and Gross, N., Die Atomanordnung des Kupferkieses und die Struktur der Berührungsflächen gesetzmässig verwachsender Kristalle: *Neues Jahrb.*, Beil. Bd. **48**, 1923, pp. 128–134.

[4] Gruner, J. W., Structural reasons for oriented intergrowths in some minerals: *Amer. Min.*, vol. **14**, 1929, pp. 230–231.

In order to study these minerals in three dimensions, and also to combine the advantages of polished section study with thin section study, an entirely new method of procedure has been adopted. This is the use of polished thick sections. A polished thick section consists of a "thin" section of about 0.5 to 1 mm. in thickness, polished on both surfaces. In making a section of this kind the following method is used: One surface of the specimen is ground and polished. This surface is firmly fastened to a glass slide by means of Canada balsam. A cut is then made with a diamond saw so as to leave about 2 mm. of the specimen attached to the slide. This is ground down to the desired thickness (usually about 0.8 mm.) and polished. The slide is next heated on a hot plate and section slipped off. The excess Canada balsam is dissolved in methyl alcohol, and the section is ready for use.

In the examination of the section both reflected and transmitted light were used. This was made possible without shifting the specimen by employing a metallographic microscope with a sub-stage attachment for transmitted light. The section was placed on a glass slide and examined. The regions in which the blebs occurred were noted on a small sketch map of the section in order that the same regions could be studied both before and after heating.

An electric furnace was set up which had a range of from 200°C. to 1000°C. Each specimen was enclosed in a pyrex glass tube of a size into which the particular specimen best fitted. A small pyrex tube was then joined onto the end of this and the container evacuated and sealed.

All specimens were rapidly cooled in a blast of air, since it was found that water quenching shattered the specimen to such an extent that it could not be used.

EXPERIMENTAL DATA ON RE-MIXING

Temperature and time are factors in any mixing or unmixing process. In order that the results of all runs made during this investigation would be consistent, the time factor was kept constant at seven days (168 hours). Runs were made at temperatures ranging from 200°C. to 500°C.

In some cases it was found possible to use the same specimen for a sequence of runs. This was most desirable since exactly the same areas were examined at the end of each run. The same specimen was used in runs 1, 2, 3 and 4. In all other runs new specimens were used each time.

Eight runs were made and these will be described separately:

Run 1.

Temperature: 200°C.

Time: 168 hours.

Results: No observable change has taken place either in the chalcopyrite blebs or in the surrounding sphalerite.

Run 2.

Temperature: 250°C.

Time: 140 hours.

Results: No change in the chalcopyrite blebs or in the surrounding sphalerite.

Run 3.

Temperature: 305°C.

Time: 169 hours.

Results: When the tube was opened, a peculiar odor was noticed which could not be identified since it was unfamiliar to those who detected it. It was certainly not hydrogen sulphide.

No observable change in either the chalcopyrite blebs or in the surrounding sphalerite.

Run 4.

Temperature: 350°C.

Time: 169 hours.

Results: The peculiar odor was again noticed, together with some hydrogen sulphide.

Although no marked change was noted, it was thought that there was a perceptible darkening of the sphalerite around each bleb of chalcopyrite. This darkening was so slight, as to be nearly unnoticeable. However, the writer felt that this indicated a proximity to the possible mixing temperature.

Run 5.

Temperature: 400°C.

Time: 168 hours.

Results: The tube was blackened on the inside and a strong odor of H_2S was noted on opening.

The specimen had apparently blown up while the run was being made, for it was in countless fragments. This was probably due to the presence of liquid inclusions. Fortunately, however, one of the fragments was found to be large enough to examine. This measured about 10 mm. by 5 mm. There were only three small areas of clear sphalerite present. One of these contained a plane of chalcopyrite blebs running through it, but they had remained apparently unchanged.

In the examination of the surface by reflected light, evidences that a mixing of chalcopyrite and sphalerite had taken place were clearly seen. There were a number of relatively large masses of chalcopyrite present as islands in the sphalerite. Surrounding each one of these, and extending perhaps over a millimeter into the sphalerite, tiny blebs of chalcopyrite had developed. These were so small as to be just barely discernable using a number 3 objective. A high power oil immersion objective was therefore used in the detailed examination of the areas. Figure 1 is a photomicrograph of such an area. The blebs showed a definite relation to the large chalcopyrite masses. The size and number of the blebs decreased in distance from the main masses, finally disappearing altogether at about 1 to 1.5 mm. from the chalcopyrite mass.

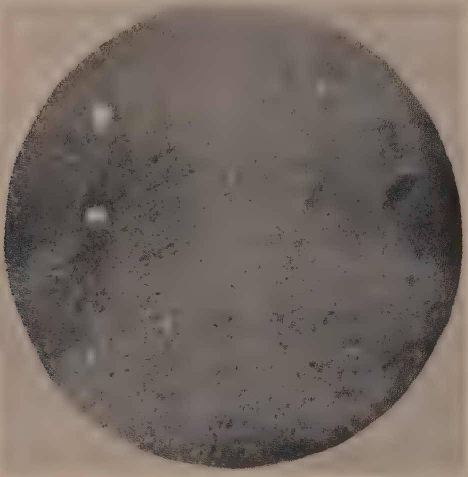

Fig. 1. Photomicrograph of an area adjacent to a large mass of chalcopyrite. Oil immersion, ×368. The large blebs are those which occur in planes throughout the sphalerite. The small blebs are those produced by mixing and unmixing as described in the text. The main chalcopyrite mass lies outside the field.

This development of minute chalcopyrite blebs in the sphalerite adjacent to large masses of chalcopyrite can only be accounted for by a mixing and unmixing process having taken place. The chalcopyrite dissolved in the adjoining sphalerite to form a solid solution of the two minerals. The cooling by air was not quite rapid enough to retain the homogeneous character of the solid solution, but allowed at least part of the chalcopyrite to exsolve, or separate out, from the sphalerite, resulting in the small blebs observed. Since the chalcopyrite in solution was more concentrated near the large mass of that mineral, more blebs have been segregated close to the mass, diminishing in number and finally disappearing further out in the sphalerite.

The fact that no mixing was apparent in the plane of blebs seen within the clear sphalerite may be accounted for by a factor not yet considered, that is the composition of the enclosing sphalerite. The only large blebs (of the type which occur in planes within the sphalerite) which may be seen by transmitted light are those in the clear portions of the sphalerite. There is no doubt that this sphalerite is of different composition than the opaque portions. The large chalcopyrite masses around which the mixing-unmixing took place are in regions of opaque sphalerite. The larger blebs of chalcopyrite, seen on the surface of the opaque sphalerite, which are undoubtedly similar to those observed in the clear portions (and are probably similarly arranged in planes within the sphalerite) also show the minute blebs of chalcopyrite concentrated about them. These facts seem to indicate that the opaque sphalerite tends to form a solid solution with chalcopyrite more readily than does the clear type. It may well be that mixing-unmixing has occurred to some extent in the clear portion, but is so slight that it cannot be observed.

It is of interest that about an island of galena in the sphalerite a similar (but less pronounced and less widespread) effect as described above for chalcopyrite, was noted. Minute blebs, probably of galena, had developed about the larger galena mass. This suggests the possibility of a solid solution of galena in sphalerite taking place. Except as a point of interest, this is not considered in the present paper.

Run 6.

Temperature: 440°C.

Time: 163 hours.

Results: The tube was blackened on the inside. Upon opening tube, there was an odor of hydrogen sulphide clearly discernible together with the unfamiliar odor noted in previous runs.

The specimen had become entirely opaque. This is taken to indicate that the mixing temperature had been exceeded with the result that some of the chalcopyrite had diffused through the sphalerite, causing discoloration. In an examination of the surface by reflected light distinct haloes were noted surrounding each chalcopyrite bleb. It is difficult to attribute this to anything other than to a difference in composition of the sphalerite due to diffusion of unmixed chalcopyrite into the sphalerite.

Run 7.
Temperature: 470°C.
Time: 168 hours.
Results: Same as for run 6.

Run 8.
Temperature: 500°C.
Time: 164 hours.
Results: Same as for run 6.

CONCLUSIONS

The results of this investigation definitely prove the existence of a solid solution of chalcopyrite in sphalerite, and the unmixing of chalcopyrite from such a solution. The unmixing temperature for this example lies within the range 350° to 400°C.

GRAPHICAL METHODS FOR THE DETERMINATION OF RETICULAR DENSITIES AND LATTICE PARAMETERS*

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Abstract

Graphical methods are known for the determination of reticular densities from a gnomonic or stereographic projection made on a plane normal to [001]. They are explained in terms of the simple shear and the polar lattice.

Our aim is to generalize these rules for the case of a projection plane normal to an arbitrarily chosen zone [*uvw*]. A general theorem is stated, covering the previous methods as particular cases. Rules for trigonal crystals referred to the Miller axes also follow directly from it. The same method yields both reticular densities and lattice parameters. A working knowledge of the method can be acquired by working out three examples, given to illustrate the rules.

I. Projection Plane Normal to [001]

According to Mallard's gnomonic theorem[1] the length of the projection-line for any face of the type ($hk1$) is proportional to the reticular area of the crystal face projected. The plane of the gnomonic projection is taken perpendicular to the zone axis [001]. In Fig. 1, for instance, the length of the projection-line P-($1\bar{1}1$) is a measure of the area of the two-dimensional unit-cell of the lattice-plane ($1\bar{1}1$). If the length of the "gnomon" (distance from the projection-point P to the plane of gnomonic projection) is equal to R, then the reticular area of the face ($1\bar{1}1$) is represented by $R/\cos\rho$, in which ρ is the angle between the gnomon and the projection-line P-($1\bar{1}1$), that is to say the *polar distance* of ($1\bar{1}1$). The reticular

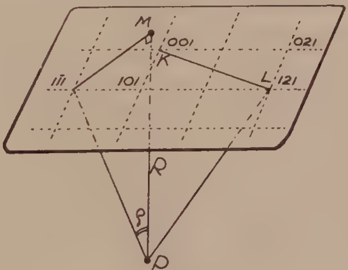

Fig. 1.

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* Translated from the Dutch by J. D. H. Donnay, Johns Hopkins University.

[1] Mallard, E.: *Traité de Cristallographie*, tome. **1**, pp. 26 and 63.

density of the face $(1\bar{1}1)$, being inversely proportional to its reticular area, is then proportional to cos ρ.

For a general face (hkl), the method is to be slightly modified. It is well known that, in a gnomonic projection, a face (hkl) is represented by $(h/l \cdot k/l \cdot 1)$. The value of cos ρ obtained for such a pole, therefore, cannot be a measure of the reticular density of the face (hkl). The true measure of that reticular density is given by the quotient $(\cos \rho)/l$.

Inspection of this expression immediately shows that another method of reticular density determination will have to be devised in the case of a face $(hk0)$, since the quotient $(\cos \rho)/l$ then leads to the indeterminate form 0/0. A consequence of Mallard's theorem is that the reticular area of the face (120), for instance, is represented by the line KL (see Fig. 1) connecting the gnomonic poles of (001) and (121), in the same way as the reticular area of the face $(1\bar{1}1)$ is represented by the line P-$(1\bar{1}1)$. In general the reticular area of any face $(hk0)$ is represented by the line obtained by joining the pole (001) with the *substituted pole* $(hk1)$.

When a crystal is given in stereographic projection, the gnomonic projection of the pole $(hk1)$ and of the pole (001) are first determined in the usual manner, then the distance between the two gnomonic poles is measured; for instance, KL in the case of the face (120), as shown in Fig. 2.

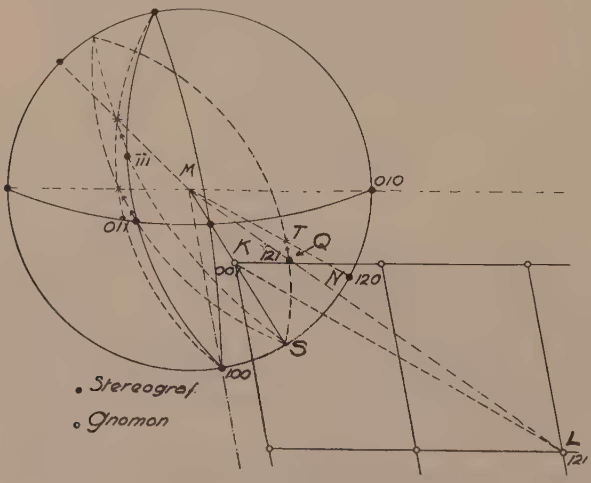

FIG. 2.

The reticular density of a face ($hk0$) can also be determined directly from the stereographic projection. The reticular area of (120) is represented by KL (Fig. 2). Let ϕ_1 be the angle NML and ϕ_2, the angle NMK. The lines KL and MN are parallel by construction. We have in the triangle MKL (Law of Sines):

$$KL = ML \frac{\sin(\phi_2 - \phi_1)}{\sin \phi_2}$$

Let R be the length of the gnomon, then:

$$ML = R \cdot \tan \rho_{121}.$$

Substituting

$$KL = R \cdot \tan \rho_{121} \frac{\sin(\phi_2 - \phi_1)}{\sin \phi_2}.$$

In the spherical triangle MSQ, the cotangent formula gives:

$$\sin(\phi_2 - \phi_1) = \frac{\tan MSQ}{\tan MQ}.$$

Likewise in the spherical triangle MST:

$$\sin \phi_2 = \frac{\tan MST}{\tan MT} = \frac{\tan MSQ}{\tan MT}$$

Substituting again:

$$KL = R \cdot \tan \rho_{121} \frac{\tan MT}{\tan MQ} = R \tan MT$$

as *arc* $MQ = \rho_{121}$.

The reticular density of the face (120) is thus proportional to *cot* MT, where the *arc* MT is obtained as follows:

(1) Join the center M of the stereographic projection to K the stereographic pole of (001), and produce to S, intersection with the primitive circle.

(2) Draw the zone circle through S and Q, the stereographic pole of (121).

(3) "Slide" Q, the stereographic pole of (121), along that zone circle until it reaches, at point T, the radius MN joining the center M to the stereographic pole N of (120).

II. Simple Shear and Polar Lattice

The significance of such a "sliding" in the stereographic projection can best be explained in terms of the properties of the *polar* *lattice*[2] (or *reciprocal lattice*). Let the dimensions of the latter be chosen in such a way that the parameter of a reticular row $[hkl]$ of the polar lattice be equal to the reticular area of the plane (hkl) of the crystal lattice. For convenience the reticular planes and rows of the *polar lattice* will be distinguished from those of the crystal lattice by means of the subscript p. Lattice point symbols will be enclosed in double brackets; $[[hkl]]_p$, for instance, represents the point of the polar lattice with "trimetric coordinates" h, k, l.

Consider now the face $(hk0)$. The corresponding row $[hk0]_p$ lies in the axial plane $(001)_p$. On this lattice-row, we find the following succession of points: $[[000]]_p$, $[[hk0]]_p$, $[[2h \cdot 2k \cdot 0]]_p$, etc. The parameter of the row $[hk0]_p$ is equal to the distance between the points $[[000]]_p$ and $[[hk0]]_p$, or (parallel lattice-planes being identical) to the distance between the points $[[001]]_p$ and $[[hk1]]_p$, which lie in ${}_1(001)_p$, the first $(001)_p$ plane[3] away from the origin (Fig. 3).

If paralle lattice-planes are numbered from the origin: ${}_1(001)_p$, ${}_2(001)_p$, ${}_3(001)_p$, etc., the above can be summarized as follows: the reticular area of a face $(hk0)$ is equal to the distance between the two points $[[001]]_p$ and $[[hk1]]_p$ of the plane ${}_1(001)_p$.

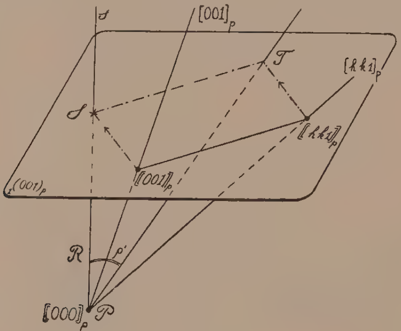

FIG. 3.

In general the foot S of the perpendicular s dropped from the origin $[[000]]_p$ on the face ${}_1(001)_p$ does not coincide with the point $[[001]]_p$. Imagine that all the parallel planes ${}_1(001)_p$, ${}_2(001)_p$, etc. (carrying with them their two-dimensional lattices), be "shoved" over each other so that the points $[[001]]_p$, $[[002]]_p$, etc., all come to lie on the perpendicular s. This movement brings the point $[[hk1]]_p$ in T. It is easy to see that the reticular area of the face $(hk0)$ is equal to $ST = R\cdot \tan \rho'$ (Fig. 3), where ρ' designates the polar distance of T.

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[2] Compare, for instance, George Tunell, *Am. Min.*, vol. **18**, p. 183, 1933.

[3] Mallard calls this plane "*plan limitrophe de l'origine*" (*Traité de Cristallographic*, tome **1**, p. 17).

In order to get a simple prescript, we can perform this "sliding" in such a manner, that it obeys the following two conditions:

(1) During the movement the lattice-row $[001]_p$ must remain in the plane defined by $[001]_p$ and the normal s to $(001)_p$.

(2) The deformation of the polar lattice will be a *homogeneous deformation*, that is to say that points lying on a straight line at equal spacings will retain these properties during the deformation.

Under these conditions, the "sliding" considered constitutes a *simple shear*[4] of the polar lattice. All the points describe rectilinear paths and their displacements are proportional to their distance from a *zero plane* (in which the displacement is of course zero). A *simple shear* is known when its zero plane, its direction, and its angular magnitude are given. The latter, in the present case, is the angle between the lattice-row $[001]_p$ and the normal s to the plane $(001)_p$.

Consider now (in the stereographic projection of Fig. 2) the normals to the faces (001), (120), (121), etc. Apply to this radiating bundle of face-normals a simple shear with the plane of the primitive circle (equatorial plane) for its zero plane and with such a direction and such a magnitude that the normal to the face (001) be brought into coincidence with the normal to the primitive circle.[5] What is the new position of the normal to the face (121)?

Notice that:

(1) During the simple shear this normal to (121) remains in the plane α defined by its original position and MS, since each point of the normal to (121) describes a path parallel to MS.

(2) The plane β defined by MN and the normal to (001) is brought by the shear to be perpendicular to the primitive circle along MN, since the normal to (001) becomes perpendicular to the primitive circle, while MN, lying in the zero plane, is not affected by the shear.

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[4] Dutch: *enkelvoudige verschuiving.* German: *einfache Schiebung.* A good example of simple shear is gliding in calcite (see Liebisch: *Grundriss der physikalischen Krystallographie,* **1896,** p. 449.

[5] In Fig. 2, the face symbol 001 has unfortunately been omitted next to the black circlet representing its stereographic projection (between $0\bar{1}1$ and 010).

The new position of the normal to (121), therefore, lies at the intersection of the two planes α and β referred to above. Its pole is stereographically projected in T (Fig. 2). The procedure to find the reticular density of the face (120) can then be stated as follows:

To the radiating bundle of face-normals, apply a simple shear that will bring the pole (001) to the center of the primitive circle. Find the polar distance ρ' of the *substituted face* (121) after the shear. The reticular density of (120) is *cot* ρ'.

III. EXAMPLE

Consider two stereographic projections of phthalylphenyl hydrazide[6] (Figs. 4 and 5). The observed forms are O, C, a, q. The fact that this substance is monoclinic will be ignored so that the choice of the axial planes will not depend on symmetry considerations. In the first projection (Fig. 4), C and a are chosen as axial planes; O_1, as unit-face. In the second projection (Fig. 5), the axial planes are q_1, q_2, and a; the unit-face is determined by assigning the symbols (101) and (011) to O_2 and O_1 respectively. The second projection can be obtained from the first by a rotation in the plane of the drawing. In both projections, the reticular densities

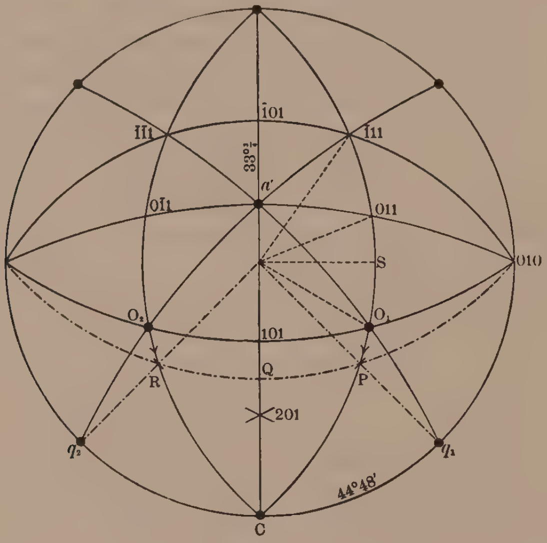

FIG. 4.

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[6] Groth: *Chemische Krystallographie*, vol. **5**, p. 168.

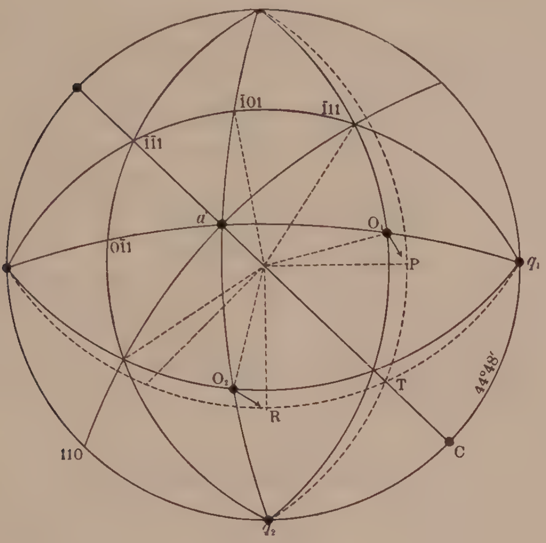

FIG. 5.

of the faces projected inside the primitive circle are determined by the formula $(\cos \rho)/l$, in which ρ is the polar distance of such a face. These reticular densities are therefore the same in both projections provided the third index l does not change.

This is no longer true for the faces projected on the primitive circle. Consider the face C. In Fig. 4, the substituted face of $C(100)$ is (101); after application of the simple shear which brings a' to the center M of the primitive circle,[7] the pole of this substituted face comes in Q.[8] Hence the reticular density of $C(100)$ is expressed by *cot arc* $MQ =$ *cot* $49\frac{1}{2}°$. In Fig. 5, on the other hand, the substituted face of $C(110)$ is (111), which is brought over to T by the simple shear,[9] hence the reticular density of $C(110)$ is given by *cot arc* $MT =$ *cot* 67°.

Table 1 gives the reticular densities of the various faces as obtained from the two stereographic projections (Figs. 4 and 5).

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[7] The letter M is unfortunately missing on the figure.

[8] In order to find this point Q we proceed in the following way: applying the rule given above, one sees that the point O_1 slides to P; therefore the circle (010)–O_1 comes in (010)–P, hence (101) comes in Q.

[9] O_1 slides to P; hence the circle q_2–O_1 comes in q_2–P; hence the new place of (111) is in the interesecting point T of the circles q_2–P and q_1–R.

Table 1
Reticular Densities for Phthalylphenyl Hydrazide

	Fig. 4			Fig. 5		
Face	Symbol	rho	Reticular density	Symbol	rho	Reticular density
c	100	49½°	cot 49½° = .854	110	67°	cot 67° = .424
q	110	59°	cot 59° = .601	010	59°	cot 59° = .601
q	$1\bar{1}0$	59°	cot 59° = .601	100	59°	cot 59° = .601
o	111	53½°	cos 53½° = .595	011	53½°	cos 53½° = .595
o	$1\bar{1}1$	53½°	cos 53½° = .595	101	53½°	cos 53½° = .595
a	001	25½°	cos 25½° = .903	001	25½°	cos 25½° = .903
	010	49°	cot 49° = .869	$\bar{1}10$	66½°	cot 66½° = .435
	101	36½°	cos 36½° = .804	112	36½°	½ cos 36½° = .402
	011	52°	cos 52° = .616	$\bar{1}12$	52°	½ cos 52° = .308

It should be kept in mind that the significance of the "reticular density" determined by the method described here is quite different from that of the "reticular density" determined by röntgenographic investigation. It is beyond the reach of geometrical crystallography to deduce the *actual reticular densities* from mere goniometric data. The only results which can be obtained are the *relative densities, after indices have been assigned to the faces.*

This means that, for each crystal, a great many space-lattices can be imagined for which the angles between reticular planes will correspond to the interfacial angles measured on the crystal.[10] All

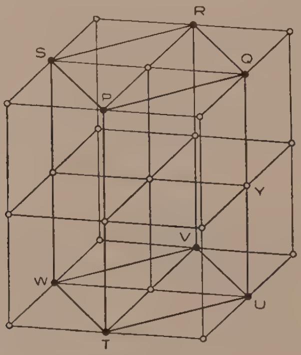

Fig. 6.

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[10] For indications as to how to make a choice between such "isogonal" lattices, see, for instance, Donnay and Mélon: Haüy-Bravais lattice . . . , *Am. Min.*, vol. **18**, p. 225, 1933.

these lattices can be derived from each other in the sense that their unit-cells are multiples of each other. Consider (Fig. 6) the space-lattice corresponding to the projection of Fig. 4. By taking the parallelopiped $PQRSTUVW$ as a new unit-cell, ignoring all other points, a new lattice can be constructed, which will be in agreement with the second projection of the crystal (Fig. 5). The second lattice is a *multiple lattice* of the first, its unit-cell being a *multiple cell* in the first lattice. Observe that all the points indicated in Fig. 6 belong to the first lattice, but that the second lattice is formed by the black dots only. So the face C (Figs. 4 and 5) or $SQUW$ in Fig. 6 has in the first lattice a reticular area $SQUW/4$, while in the second lattice the reticular area is $SQUW$. In the same manner we find for the face q_1 or $PQUT$ in Fig. 6 the reticular areas in the first and second lattices respectively $PQUT/2$ and $PQUT$; so O_1 or SRY. Therefore the reticular densities of the faces C, q, O, etc., referred to the second lattice, can be obtained by dividing their densities in the first lattice by 4, 2, 2, etc. Again, the numbers thus found can only be taken to indicate the *ratios* between the reticular densities of the faces in question. The unit chosen to express these reticular densities may be changed at will. If, for instance, we wish the reticular density of the face q to be the same in both lattices,[11] we divide the edges of the unit parallelopiped $PQRSTUVW$ by $\sqrt{2}$. This was done to arrive at the numbers listed in Table 1.

IV. Projection Plane Normal to [*uvw*]. General Theorem

The preceding sections are limited to the case where the plane of projection is normal to the zone [001]. Rules for the determination of reticular densities in the case of a projection plane perpendicular to any zone [*uvw*] will now be derived. The trend of the reasoning will be similar to that of Section II.

A general theorem can be stated as follows:

When a crystal is projected on a plane normal to an arbitrarily chosen zone [*uvw*],

(1) the reticular density of a face (*hkl*) which does not lie in the zone [*uvw*] is given by the formula:

---

[11] The absolute values of the parameters of the "lattices" considered in this paper are not known since they are derived from angular magnitudes only; this is why it has been suggested to call such lattices "relative" or "elastic" in contradistinction with the "absolute" lattice as determined by x-ray investigation.

$$\frac{\cos\,\rho}{hu+kv+lw},$$

IN WHICH ρ IS THE POLAR DISTANCE OF THE FACE (hkl);

(2) THE RETICULAR DENSITY OF A FACE (pqr) WHICH DOES LIE IN THE ZONE $[uvw]$ IS GIVEN BY THE FORMULA:

$$\frac{\cot\,\rho'}{u^2+v^2+w^2},$$

WHERE ρ' IS THE POLAR DISTANCE OF AN AUXILIARY FACE $(p+u\cdot q+v\cdot r+w)$, AFTER APPLICATION OF A SIMPLE SHEAR IF THE POLE OF THE ZONE AXIS $[uvw]$ DOES NOT COINCIDE WITH THE POLE OF THE FACE (uvw).

The demonstration of the first half of this theorem is based on the following property of the polar lattice. The reticular area of the face (hkl) is equal to the distance between the points $[[000]]_p$ and $[[hkl]]_p$. Let d be that spacing. Consider the series of parallel planes ${}_1(uvw)_p$, ${}_2(uvw)_p$, etc., of the polar lattice. One of these planes, ${}_n(uvw)_p$, passes through the point $[[hkl]]_p$. It is well known that its *serial number* [12] is equal to:

$$n=hu+kv+lw.$$

Now if the plane ${}_1(uvw)_p$ is chosen as plane of projection,[13] the length of the projection line of the face (hkl) is equal to $d/(hu+kv+lw)$. This length, however, is also equal to $R/\cos\,\rho$, where R is the length of the gnomon and ρ is the polar distance of (hkl). We may therefore write

$$d=\frac{R(hu+kv+lw)}{\cos\,\rho}.$$

In other words, the reticular density of (hkl) is proportional to $(\cos\,\rho)/(hu+kv+lw)$.

For a face (pqr) which lies in the zone $[uvw]$, this formula cannot be applied, since it leads to the indeterminate form 0/0. The reason for this is that the line connecting the points $[[000]]_p$ and $[[pqr]]_p$ lies in the $(uvw)_p$ plane passed through the origin. Instead of using these two points, it is more convenient to consider the points

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[12] Dutch: *volg nummer.*

[13] According to the usual practice. See E. Mallard: *Traité de Cristallographie,* vol. **1,** pp. 26 and 63.

$[[uvw]]_p$ and $[[u+p\cdot v+q\cdot w+r]]_p$. Their spacing d' is equal to that of the former two points, and they are lying in the plane $_m(uvw)_p$, with a serial number equal to

$$m=u^2+v^2+w^2.$$

Taking the plane $_1(uvw)_p$ as zero plane, apply a simple shear which will bring the point $[[uvw]]_p$ on the normal to the face $(uvw)_p$. Call ρ' the polar distance of $[u+p\cdot v+q\cdot w+r]_p$ after the shear. The distance d' gives the reticular area of (pqr). It is m times larger than R *tan* ρ', that is to say

$$d'=(u^2+v^2+w^2)R \tan \rho'.$$

In other words, the reticular density of (pqr) is proportional to $(cot\ \rho)'/(u^2+v^2+w^2)$.

V. PARTICULAR CASES OF THE GENERAL THEOREM

The formulae of Section I can be derived from the general theorem as particular cases by letting u, v, w, equal 0, 0, 1, respectively. The formulae obtained in that manner hold good for the ordinary projections of isometric, hexagonal, tetragonal, orthorhombic, monoclinic, and triclinic crystals. No simple shear is needed in the case of isometric, hexagonal, tetragonal, or orthorhombic crystals. The same advantage can be obtained for monoclinic crystals if they are projected on (010); in that case, the reticular density of a face (hkl) is found by dividing *cos* ρ by k, not by l.

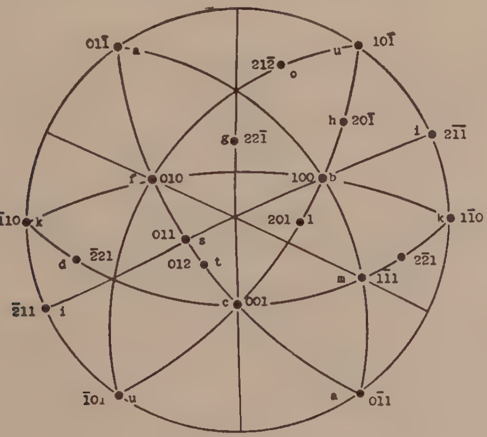

FIG. 7.

The determination of reticular densities is somewhat more intricate for *trigonal crystals* referred to the Miller system of coördinate axes. It follows from the general theorem for $u=v=w=1$. For any face (hkl) which does not lie in the vertical zone, the reticular density is then $(\cos\rho)/(h+k+l)$, in which ρ is the polar distance of (hkl). For a vertical face (pqr), the reticular density is equal to $\frac{1}{3}\cot\rho'$, where ρ' is the polar distance of the substituted face $(p+1\cdot q+1\cdot r+1)$.

For pseudo trigonal crystals described with reference to Miller axes, the above formulae are also applicable provided the necessary simple shear be applied before reading ρ' for the faces of the vertical zone.

Tarbuttite provides a good example of the last case. Guided by the *complex symbol* found by Fedorov for that mineral, we have departed from Spencer[14] in adopting the pseudotrigonal orientation projected in Fig. 7. The letters used to designate the crystal forms are those of Spencer's original description. The Fedorov orientation can be obtained from that of Spencer by the transformation $010/\bar{1}00/101$. The reticular densities are listed in Table 2.

TABLE 2

RETICULAR DENSITIES OF TARBUTTITE

Face	Symbol	Substituted face	Reticular density
c	001	—	cos 44° = .719
b	100	—	cos 49° = .656
f	010	—	cos 50° = .643
a	$0\bar{1}1$	102	$\frac{1}{3}$ cot $30\frac{1}{2}$° = .566
u	$\bar{1}01$	012	$\frac{1}{3}$ cot 31° = .554
k	$\bar{1}10$	021	$\frac{1}{3}$ cot 34° = .494
s	011	—	$\frac{1}{2}$ cos 30° = .433
m	$1\bar{1}1$	—	cos $66\frac{1}{2}$° = .399
g	$12\bar{1}$	—	$\frac{1}{2}$ cos 46° = .348
—	111	—	$\frac{1}{3}$ cos 4° = .333
i	$\bar{2}11$	$\bar{1}22$	$\frac{1}{3}$ cot 48° = .300
t	012	—	$\frac{1}{3}$ cos 30° = .289
l	201	—	$\frac{1}{3}$ cos $33\frac{1}{2}$° = .278
o	$21\bar{2}$	—	cos 75° = .259
d	$\bar{2}21$	—	cos $76\frac{1}{2}$° = .233
e	$2\bar{2}1$	—	cos 77° = .225

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[14] Spencer, L. J., *Min. Mag.*, vol. **15**, p. 22, 1908.

Remember that, for instance, the angle $\rho' = 30\frac{1}{2}°$ for the substituted face (102) is read on the stereographic projection after the pole (111) has been brought to the center of the primitive circle by the proper *simple shear* (here $3\frac{1}{2}°$).

VI. CENTERED LATTICES

The preceding sections all have reference to crystals which can be described by means of *simple lattices*. If one is dealing with a *centered lattice*, the reticular densities are first determined as if the lattice were simple, then corrected as follows:

(1) For a *body-centered lattice*, the reticular densities of the faces for which $(h+k+l)$ is even are multiplied by two.

(2) For an *all-face-centered lattice*, the reticular densities of the faces with all indices odd are multiplied by two.

(3) For a lattice *centered in one face xy only*, the reticular densities of the faces for which $(h+k)$ is even are multiplied by two.

VII. DETERMINATION OF PARAMETERS

The gist of the method lies in the following remarks:

(1) The reticular density of a plane is the reciprocal of its reticular area; the linear density of a row is the reciprocal of its parameter.

(2) (a) The determination of the reticular areas of a crystal K_1 resolves itself into the determination of the parameters of its polar crystal K_2.

(b) The determination of the reticular densities of a crystal K_1 resolves itself into the determination of the linear densities of its polar crystal K_2.

Inversely, since the crystal K_1 is the polar crystal of K_2 (just as K_2 is the polar of K_1), the determination of the reticular densities of K_2 is equivalent to the determination of the reciprocals of the parameters of K_1. Hence in order to find the parameters of K_1 graphically, the first step is to derive the stereographic projection of K_2 from the given stereographic projection of K_1. The face poles of K_1 are the points where the edges of K_2 intersect the sphere of projection, or, in other words, *the stereographic projection of K_2 is at the same time the cyclographic projection of K_1*. Hence, the rules of reticular density determination, applied to the cyclographic projection of a crystal, yield the reciprocals of the parameters of that crystal.

The stereographic projection of a crystal K_1 being given, the corresponding cyclographic projection is found by plotting the pole of each zone circle (this is easily done by means of a Wulff net) and then by rotating the projection until the polar crystal K_2 is projected on a face of the given crystal K_1.

EXAMPLE.—Starting from the stereographic projection of sodium molybdo-tellurate[15], the corresponding cyclographic projection (Fig. 8) is derived in that manner. The values of some of the parameters, graphically determined on the projection are listed in Table 3.

The order of accuracy of the graphical solution may be judged by

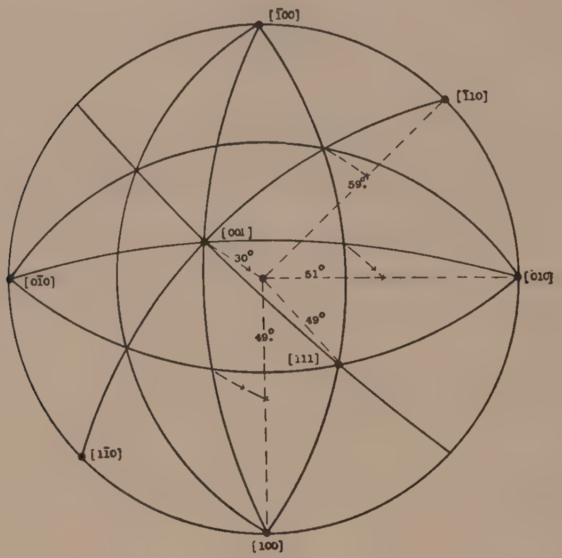

FIG. 8.

TABLE 3
PARAMETERS OF SODIUM MOLYBDO-TELLURATE

Lattice row	Parameter	Parameter ($b=1$)
[100]	tan $49\frac{1}{2}$° = 1.171	0.948
[010]	tan 51° = 1.235	1.000
[001]	sec 30° = 1.156	0.940
[110]	tan $59\frac{1}{2}$° = 1.698	1.234
[111]	sec 49° = 1.524	1.375

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[15] Donnay and Mélon: *Am. Min.*, vol. **18**, p. 243, Fig. 6, 1933.

comparison of these results with the computed values on record:

$$\breve{a}:\bar{b}:\acute{c}=0.9548:1:0.9344.$$

In this case, the error is less than one *per cent.*

The above method for the determination of parameters only applies to simple lattices. In the case of a *body-centered lattice*, the parameters found for rows with all three indices odd must be divided by two. If the lattice is of the *face-centered* type, the parameters of the rows with two indices odd should be divided by two. For a lattice centered in one face xy only, the parameters of the rows for which $(h+k)$ and l are even should be divided by two.

The graphical determination of reticular densities and parameters described in this paper is sufficiently accurate for a number of purposes when made on the usual Wulff net (10 cm. radius). For greater accuracy the Schiebold stereographic net (50 cm. radius), sold by the firm R. Seifert in Hamburg, may be found serviceable.

ACKNOWLEDGMENTS

Our sincere thanks are due to Dr. J. D. H. Donnay for the translation of this paper, the idea of which largely grew out of the conversations carried on with him during the Summer of 1933. We owe Figs. 4 and 5 to the courtesy of Dr. J. H. Haan; they are reprinted from his *Kristallometrische Determineeringsmethoden.*

X-RAY STUDY OF NARSARSUKITE, $Na_2(Ti, Fe)Si_4O_{11}$

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Massachusetts Institute of Technology, Cambridge, Mass.

Summary

Narsarsukite $Na_2(Ti, Fe)Si_4O_{11}$, a tetragonal crystal, has been found to have cell dimensions "a" = 10.74Å, "c" = 7.90Å, and to contain 4 molecules per unit cell. The axial ratio is 0.735 instead of 0.5235 (Dana), and the perfect cleavage assumed by Dana to be $\{110\}$ becomes $\{100\}$ in terms of the new axes. The space group is one of the three possibilities S_4^2, C_4^5, C_{4h}^5, but due to complete lack of knowledge of the crystal class it has not been possible to decide between these, nor to make a complete structure determination.

Introduction

Narsarsukite is a rather rare mineral found in Greenland. According to Dana, it is tetragonal with $c = 0.5235$, prominent $\{110\}$ and $\{001\}$ cleavages, found commonly as prisms. The specific gravity is given as 2.751 and the analysis as follows:

SiO_2 = 61.63
Al_2O_3 = 0.28
Fe_2O_3 = 6.30
TiO_2 = 14.00
MnO = 0.47
MgO = 0.24
Na_2O = 16.12
F = 0.71
H_2O = 0.29
100.04

Unit cell

The crystals used in the present investigation were cleavage fragments[1] measuring about 1×1.5×2 mm. in size with what, according to Dana, were perfect $\{110\}$ faces. Rotation photographs (Mo $K\alpha$) were made by rotating about the 001, 110 and 100 axes. The following primitive translations were found:

$$a_1(001) = 7.90\text{Å}$$
$$a_2(110) = 10.74\text{Å}$$
$$a_3(100) = 15.19\text{Å}$$

---

* New York State College of Ceramics, Alfred University. Work done at M.I.T.

[1] We are indebted to Prof. C. Palache of Harvard University for the crystals used, and to Mr. H. Berman for selecting and examining them for us.

Since $a_3 = a_2\sqrt{2}$, the true unit cell has the dimensions:

$$c = 7.90\text{Å}$$
$$a = 10.74\text{Å}$$

The cleavage faces called {110} by Dana then become in reality the {100} and {010} faces. The axial ratio 0.5235 corresponds to 7.90/15.19 = 0.520. However the true axial ratio is $c/a = 7.90/10.74 = 0.735$.

FORMULA AND NUMBER OF MOLECULES PER CELL

Since Dana gives no formula, and from the analysis alone it is somewhat difficult to arrive at a satisfactory one, the number of molecules per cell on the basis of a molecular weight of 100 was calculated.

$$n = \frac{V\rho N}{M} = \frac{912\times10^{-24}\times2.751\times0.606\times10^{24}}{100} = 15.25$$

V = vol. of cell in cc.
ρ = density
N = Avogadro number
M = molecular weight

By multiplying the number of molecules of each oxide contained in a weight of 100 by this number, the number of molecules of each oxide per cell was arrived at. The cations and anions per cell could then be calculated. In this manner it was found that, if MgO, MnO, Al_2O_3, Fe_2O_3 and TiO_2 were grouped together, the cell contained approximately 16 Si atoms, 4 Ti and others, 8 Na, and 44 O, corresponding to a cell containing 4 molecules of the formula $Na_2TiSi_4O_{11}$, with the titanium partially replaced by appreciable quantities of iron and negligible amounts of magnesium, manganese, and aluminum. The fact that the number of atoms of each kind comes out almost to an exact multiple of 4, a number highly appropriate to a tetragonal crystal, makes the deduced formula very reasonable.

SPACE GROUP

A series of 15° oscillation photographs about the c axis were taken and indices assigned to about 300 reflections. It was found that no reflections appeared for which the sums of $h+k+l$ were odd.

Because of the fibrous nature of the cleavage it was found impossible to produce a satisfactory sample for a Laue pattern parallel to {001}, but a Laue pattern parallel to {100} showed no plane of symmetry parallel to the tetragonal axis. From the above data the space group is limited to one of the three possibilities S_4^2, C_4^5, C_{4h}^5. Because of complete lack of knowledge of the crystal class, it was not possible to decide between these three possible space groups, nor to make a complete structure determination.[2]

[2] Since completion of this work it has come to our attention that a similar study by Gossner and Strunz, *Zeit. Krist.*, Vol. **82**, p. 151, 1932, leads to essentially the same results.

NOTES AND NEWS

VIRGINIA STAUROLITES AS GEMS

Joseph K. Roberts, *University of Virginia.*

In Virginia well over a quarter of a century ago people began wearing staurolite crystals as watch charms, and pendants for necklaces. This custom has grown until at the present time nearly every town in the state has one or more places where the staurolites are for sale, some of which are genuine and some artificial. During the summers of 1922 and 1923, the writer visited the better localities for collecting in Henry and Patrick Counties, Virginia. In the summer of 1916 a collection was made at Ducktown, Tennessee, and in 1917 a collection in Fannin County, Georgia. Being somewhat familiar with the mineral in its occurrence, and after seeing so many on the market, and some of the ways they are prepared, the purpose of this brief article is to call attention to the Virginia localities, and some of the ways the artificial stone is made and sold. Practically all of the socalled staurolites sold as gems are of the right angle pattern. In Henry and Patrick Counties this form of twinning is rare compared to others, and since the cross of about 90 degrees is sold almost exclusively, the writer offers an explanation.

The main locality in Virginia extends from near the city of Lynchburg southwestwardly into North Carolina, the better portion for collecting being in Henry and Patrick Counties. This belt lies in the portion of Virginia known as the Piedmont province, one of the natural divisions of the state. Its extent in Virginia is approximately 60 miles with a maximum width of about 10 miles in the vicinity of Sanville, near the Patrick-Henry County line.

The rock series in which the staurolites occur is known as the Wissahickon schist according to the latest issue of the state geological map, and on the older map as the crystalline schists and gneisses. The staurolite-bearing schist is a light to dark gray color, composed of tabular minerals for the matrix with the staurolites scattered through the rock. The schist weathers and becomes colored by hydrous iron oxides, leaving the crystals upon and within the soil, where they may be collected in large numbers. Collecting from the fresh rock is slow and difficult as the schist adheres to the crystals, and removing it with a knife or file, the staurolites generally break. The crystals show various degrees of weathering, and practically all of them contain small cavities.

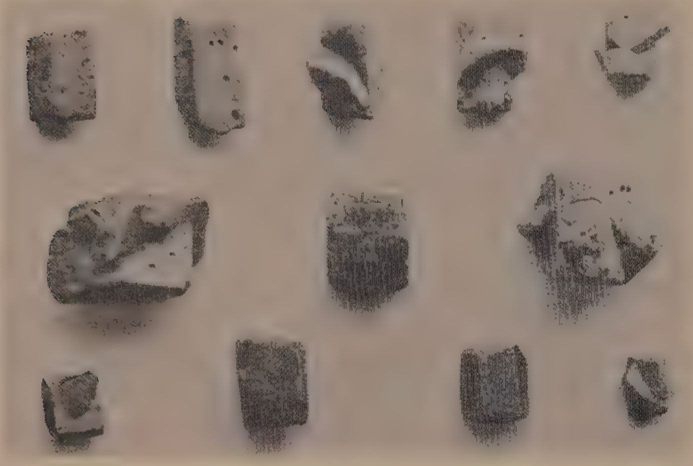

FIG. 1. Crystals of staurolite weathered from the staurolitic schist, Henry and Patrick Counties, Virginia. (Reduced $\frac{1}{3}$.)

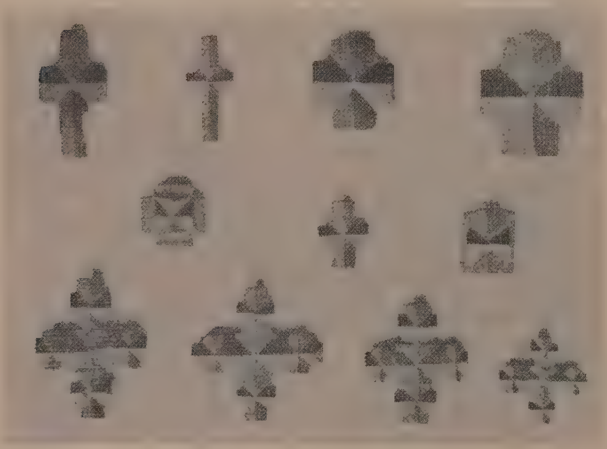

FIG. 2. Staurolites or Fairy Stones as sold on the market; upper row and the middle specimen of the middle row may be made from the natural crystals; bottom row are artificial. (Reduced $\frac{1}{3}$.)

The crystals occur as simple prisms and in some of the known forms of twinning, but rarely with good terminations. The simple or untwinned crystals occur in subordinate numbers compared to the twinned forms. The majority of specimens collected are twinned on the pyramid, the individuals penetrating each other at an angle of approximately 60 degrees. Quite a few twinned forms occur in which the twinning is on the plane $x(032)$, thus making a cross of almost 90 degrees.[1] Very few good specimens, known as triplets, were found; these result from combined twinning on the $x(032)$ and $z(232)$ planes. No crystals twinning on the prism or after the hemimorphic pattern were noted. Specimens twinning on the pyramid z (232) are commonly known as the "Cross of St. Andrew," and those on the brachydome $x(032)$ as the "Roman Cross." The latter pattern is the favorite one for use as gems, and the large numbers on the market compared with their relative scarcity naturally leads the collector to a suspicion.

In regard to the staurolites found on the market, it is quite possible that many of these of the simple Roman pattern are genuine, having been filed and smoothed from the natural twinned specimens. Such specimens are shown in Fig. 2 in the upper row. Many, however, are purely artificial, as the end specimens of the middle row and the bottom row in Fig. 2. On finding some of the natives making such specimens as shown on the bottom row of Fig. 2 leads any one to doubt all of the very smooth specimens.

The artificial crosses are cut from a talcose schist of a gray color and of a hardness ranging from about 2½ to 3. Thin and narrow band saws fitted on modified grooves of sewing machine pulleys, are used to cut the crosses into rough form; the finishing is done by flat files. The final step is to soak the cross in linseed or some other oil, which gives a dark brown color resembling the color of a natural staurolite when carried in a pocket for some time. Even the compound crosses have been taken seriously by some mineralogists, but the straw which broke the camel's back was that some of the natives produced a staurolite cut after the pattern of a swastika.

The descriptive matter used in advertising these gems is most ludicrous. These "fairy stones" are reputed to bring good fortune, hence their other common name, "lucky stones." Their origin is attributed to the crystallized tears shed at the time of the crucifixion, or to stars falling from Heaven on this occasion. In mounting the gems jewelers often place a gold tip on one prong, sometimes on opposite prongs, and frequently on all prongs. The natural crystals on weathering develop small cavities, and this serves to identify them at times, though holes are bored into the artificial stones to make them look genuine. Some of the advertisements venture statements that this gem is found only in Virginia, and if the reference is made only for the artificial ones, such a statement may be sustained. Even some popular articles have advertised the Virginia staurolites and probably were sincere in so doing. Most all texts on mineralogy now make mention of Patrick County, Virginia, but very few mention Henry County, and the latter is really better in many respects for good collecting.

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[1] The writer wishes to express appreciation to his colleague, Dr. Arthur A. Pegau in checking the twinning of the crystals.

UNUSUAL CRYSTAL HABIT OF CASSITERITE

John W. Gruner, *University of Minnesota*

The writer is indebted to Mr. A. J. Haley of Pulacayo, Bolivia, for the loan of a crystal group ($3'' \times 2'' \times 1''$) of cassiterite which shows very unusual crystal habit. The specimen is from Araca, Bolivia. As a whole it is reddish brown in appearance, due to a coating of iron oxides on portions of the prism faces and between the crystals. A few small iron stained crystals of quartz are in the group. The individual cassiterite crystals, on an average, are about 2 to 4 millimeters square and 6 to 12 millimeters long. They are brownish yellow in color and translucent.

The unusual features about them are that only first order prisms and first order unit pyramids occur, and that the prism faces are about 2 to 3 times as long as they are wide. All the prism faces are peculiarly furrowed parallel to the elongation, suggesting curved vicinal faces with very large intercepts along the $\acute{c}$ axis. The curving is convex with respect to the nearest prism edges resulting in divergence of the furrows toward the ends of the prism. No doubly terminated crystals or twins, can be found in the group. Measurements on the well developed pyramids show angles $111 \wedge \bar{1}\bar{1}1 = 87°02'$. Therefore $111 \wedge 110 = 46°29'$, which approaches very closely the theoretical value of $46°27'$.

BARKER'S DETERMINATIVE ANGLES FOR CASTANITE

J. H. D. Donnay, *Johns Hopkins University.*

The mineral castanite provides an interesting example in the search for its classification angles according to the Barker method.[1]

Castanite, the validity of which as a definite mineral was established by A. F. Rogers[2] in 1930, crystallizes in small triclinic crystals. Six forms were observed by Rogers, who assigned to them the following symbols: *b*(010), *M*($1\bar{1}0$), *m*(110), *c*(001), *e*(011), and *s*(112). This crystal does not possess a four-plane zone, but exhibits four three-plane zones. The zone *msc* offers the greatest angle of all ($cm'=90°50'$), it is therefore completed harmonically by inserting a fourth pole I between *c* and *m'*. The stereographic projection is now rotated so that the newly created four-plane zone comes into vertical position. Note that the completion of this zone does *not* simultaneously create another four-plane zone; neither does it prevent the application of the principle of simplest indices. We are now dealing with a case of "three terminal planes" in triangular arrangement. Zones passed through terminals taken two at a time meet the primitive circle in observed faces with simplest indices. Of these three zones, the zone *ceb* offers the largest angle ($b'c=89°36'$); it is harmonically completed by the insertion of a pole II between *b* and *c'*. The problem is now one of "four terminals in the *T*-formation." Note that the completion of the zone *ceb* does not lead to complex indices.

The two four-plane zones now available intersect in *c*. The zone *bmM*, meeting the primitive circle in *m*, the pole next but one to *c*, is completed harmonically by the addition of a pole III inserted in the largest angle ($M'm'=71°37'$). No ambiguity exists at this stage. The central quadrilateral is built up from the five terminals so as to comply with the principle of simplest indices. The planes *M*, *m*, and *c* are given pinacoidal symbols, they are to be the so-called "cubic" faces. The "dodecahedral" faces will be *s* and I on the primitive circle, *b* and III on the central quadrilateral. The planes *e* and II are taken for the "octahedral" planes. All the observed forms are thus given simplest notation. The axial and parametral planes are unambiguously determined. In other words, the unit-cell of the lattice adopted for determinative purposes is determined in shape. What remains to be done is the naming of the axes of reference and orienting them.

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[1] Barker, T. V., *Systematic Crystallography*, London, **1930.**

[2] Rogers, A. F., Castanite, a basic ferric sulfate from Knoxville, California, *Am. Mineral.*, vol. **16,** pp. 396–404, 1931. Attention is called to two misprints in the paper: on p. 397, in the explanation of the figures, read *M*{$1\bar{1}0$} instead of *m*{$1\bar{1}0$}; same page, second interfacial angle listed, read (010)∧($\bar{1}10$) instead of (010)∧(110).

By applying the hypomonoclinic principle, the plane *c* must be chosen as the side pinacoid[3] **b**{010}. The condition that **cr**(001:101) must be less than **ra**(101:100) determines the notation of the forms *M* and *m*, which become **c**{001} and **a**{100}, respectively. Now the face **c**(001) must slope forward and to the right, hence the faces *m*(110), *c*(001), and *M*($1\bar{1}0$) become **a**(100), **b**(010), and **c**(001), respectively.

The Barker classification angles must now be computed from the values given by Rogers in the original description. The calculations are somewhat delicate on account of certain angles being very close to 90°. The results are embodied in the "filing card" of castanite according to the Barker scheme.

CASTANITE, $Fe_2O_3 \cdot 2SO_3 \cdot 8H_2O$ (Rogers, A. F., *Am. Mineral.*, vol. **16**, p. 396, 1931).

ELEMENTS: $a:b:c = 0.726:1:0.895$.

$\alpha = 89°50'$, $\beta = 91°10'$, $\gamma = 78°46'$.

TRANSFORMATION: $110/001/1\bar{1}0$.

FORMS	*Old:*	*b*(010),	*M*($1\bar{1}0$),	*m*(110),	*c*(001),	*e*(011),	and *s*(112).
	New:	$10\bar{1}$,	001,	100,	010,	$11\bar{1}$,	110.

ANGLES: cr = 31°57′, ra = 39°40′, **am = 54°36′**,
mb = 34°34′, bq = 39°28′, qc = 49°24′.

The reader who desires to reproduce the above derivation by means of a stereographic net will obtain sufficiently accurate values of the classification angles (by reading them on the projection) to check the correctness of his adopted setting. The coördinate angles for the forms of castanite have been computed by Rogers from his interfacial angles and can be found on page 401 of his paper. The interfacial angles measured by Rogers are as follows: *bm* = 61°34′, *bM′* = 46°49′, *ce* = 42°32′, *eb* = 47°52′, and *cm* = 89°10′.

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[3] The letters referring to the new setting (Barker) are printed in bold face.

BOOK REVIEW

THE GEOLOGY AND ORE DEPOSITS OF SIERRA COUNTY, NEW MEXICO. GEORGE TOWNSEND HARLEY. Bulletin No. 10 of the State Bureau of Mines and Mineral Resources of the New Mexico School of Mines. Socorro, New Mexico. Price 60¢.

While Sierra County is one of the smaller counties of New Mexico it has been its third largest producer of mineral wealth. The production of ores, concentrates and bullion has totaled approximately $21,000,000. Silver and gold being the chief metals mined. The period of greatest mining activity was from 1877 to 1901, but with the recent rise in the price of gold interest in the ore deposits of the county has increased greatly, and mining and prospecting have had a marked revival.

The bulletin has been prepared primarily for the use of prospectors and operators of mining property in Sierra County, and for non-residents interested in the development of the resources of the region. The first part of the report includes a description of the rocks and geological history of the county, a discussion of the ore bodies and a brief review of some practical points involved in the search for and economical exploitation of the ore. The second part is devoted to a detailed description of the vein systems and other modes of occurrence of the ore in each of the mining districts within the county. The bulletin consists of 220 pages and 30 illustrations, including a detailed geologic and topographic map of the Hillsboro district.

W. F. H.

NEW MINERAL NAMES

Ferruccite

GUIDO CAROBBI: Sulla presenza di un nuova minerale fra i prodotti dell'attivitá fumarolica del Vesuvio, *Periodico di Mineral., Roma*, vol. **4,** No. 3, pp. 410–422, 1 pl., 2 text figs.

NAME: In honor of Ferruccio Zambonini.

CHEMICAL PROPERTIES: A sodium fluoborate $NaBF_4$. Soluble in water. An analysis is given of the water soluble and insoluble portions of the mineral mixture.

CRYSTALLOGRAPHICAL PROPERTIES: (On artificial crystals) Orthorhombic, with b, a, c, m. Habits, Tabular to b and to c. $a:b=0.9169:1$. $a:m=42°31'$.

OPTICAL AND PHYSICAL PROPERTIES: (On artificial crystals) Biaxial, positive, $2V=11°25'$. $\alpha=1.301$, $\beta=1.3012$, $\gamma=1.3068$. $X=c$, $Y=b$, $Z=a$.

Plane of optic axes=(010), Sp. Gr. 2.496.

OCCURRENCE: Found with other fluoborates and fluosilicates as a fumarolic product at Vesuvius. Fractional crystallization gave (1) Insoluble 71%, (Fluosilicate of aluminum and iron), (2) K_2SiF_6, (3) potassium fluoborate, (4) boric acid and sodium fluosilicate, (5) sodium fluoborate.

W. F. F.

Glaucocerinite

E. Dittler and R. Koechlin: Über Glaukokerinit, ein neues Mineral von Laurion (Glaucocerinite, a new Mineral from Laurion). *Centr. Min. Geol. und Paleont.*, Abt. A, No. 1, pp. 13–17, 1932.

Name: From the Greek words for blue and waxy.

Chemical Properties: A hydrous basic sulfate of zinc, aluminum and copper: $Zn_{13}Al_8Cu_7(SO_4)_2O_{30}\cdot 34H_2O$. Analysis: ZnO 37.95, Al_2O_3 15.40, CuO 19.26, SO_3 5.79, H_2O+ 5.69, H_2O- 16.31; Sum 100.40. Soluble in hydrochloric acid.

Physical and Optical Properties: Color sky blue to turquois blue, at times greenish, and gray or brownish through impurities. Soft, waxy to radial fibrous. Sp. Gr. 2.749.

Extinction parallel. $\gamma = 1.542 \pm .001$. Birefringence is high.

Occurrence: Found as a fine botryoidal coating on adamite or with gypsum in vugs in smithsonite at Laurion.

Discussion: Differs from zinkaluminite chiefly in its content of copper.

W. F. F.

Discredited Species

Zamboninite

A. Ferrari and R. Curti: Ricerche sopra l'eventuale esistenza del minerale zamboninite, *Periodico di Mineral., Roma*, vol. **4,** No. 3, pp. 464–472, 1 pl.

This mineral (Cf. *Am. Mineral.*, vol. **15,** p. 275, 1930) is shown by *x*-ray examination to be a mixture of fluorite and sellaite.

W. F. F.